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This paper is accompanied by a colored geologic map. Price 35 cents, prepaid. Remittance should accompany order.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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CRUSTAL LAYERS OF THE CONTINENTS AND OCEANS

By Beno Gutenberg

Abstract

It has been assumed heretofore that the relatively large phase called $\bar{P}$ in records of near-by earthquakes in the continents is the direct longitudinal wave, and that the corresponding velocity of about 5.6 km/sec is characteristic of the "granitic layer." This leads to contradiction with the calculated origin-time of transverse waves in earthquakes and cannot be reconciled with observations of the longitudinal waves from blasts. It seems more likely that the velocity of longitudinal waves below the sediments is about 6 km/sec, increases to 6½ km/sec or more at a depth of roughly 10 km, and possibly decreases below a depth of between 10 and 15 km. Such a decrease in velocity is to be expected in rocks with an appreciable content of quartz, since in laboratory experiments a decrease of elastic constants of quartz with increasing temperature has been found approaching the temperature at which transformation from alpha- to beta-quartz occurs (corresponding to a depth of 25 km or more). At the bottom of a deeper layer with higher velocity (usually 7–7½ km/sec) the Mohorovičić discontinuity at a depth of between 30 and 40 km, but deeper under some mountain chains, forms the boundary between the simatic crustal layers and the ultra-basic material below (wave velocity 8.2 km/sec). Most earthquake foci seem to be in the hypothetical low-velocity layer.

Geophysical and geological evidence indicates a greater difference between the structure of the Pacific basin and the surrounding continental areas than between the bottom of the Atlantic or Indian oceans and the surrounding shelves or continents. In the Pacific, the surface layers seem to consist of sediments, erupted and perhaps some simatic material. Below them is probably ultra-basic material with a boundary (Mohorovičić discontinuity) at a depth of only a few kilometers. In the Atlantic Ocean (and probably similarly the Indian Ocean) the transition from the continents to the basins seems to be more gradual, and the Mohorovičić discontinuity seems to be at greater depth than in the Pacific but much shallower than in the continents. While at present there is no indication of extensive sialic material in the bottom of the Pacific, there may be limited areas with sialic material at least in the eastern part of the Atlantic Ocean while relatively basic simatic (but not ultra-basic) material seems to be close to the surface at least in parts of the western Atlantic basin.

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The first definite results concerning the velocity of propagation of earthquake waves in the upper parts of the continental crust were published by A. Mohorovičić (1910). He found a velocity of about 5.6 km/sec for the longitudinal waves ($\bar{P}$) in an upper layer (later frequently called "granitic layer") and a velocity of about 7.8 km/sec beginning at a depth of about 50 km (P_n). Data from other regions furnished similar results with one or two addi-

tional layers in some instances. (For a summary, *see* Gutenberg, 1943a.) In most instances the velocity found for the waves through the upper most layer (not including the sediments) was between about 5.5 and 5.7 km/sec. These waves are very strong. (See examples of seismograms in Gutenberg, 1943a.)

In 1921, a large explosion occurred at the chemical works at Oppau (near Mannheim, Germany). The explosion was recorded at 10 stations between about 20 and 370 km from the source. Hecker (1922) found a velocity for the first longitudinal wave of 5.73±0.04 km/sec; others calucated velocities between 5.4 and 5.6 km/sec. No definite record was obtained of waves passing through deeper layers, although such a wave may have been recorded at Göttingen.

Differences between velocities found from earthquakes and those from artificial explosions were first noticed when Wiechert (1926) investigated records of a number of artificial explosions recorded by highly sensitive instruments at Göttingen. One of these instruments had a mangification of more than one million. He found that in a relatively large part of northern Germany the velocity in a layer beginning at a depth of about 2 km is 5.98 km/sec. Stille suggested that this may be the velocity in the Variscan bed-rock. Again, the distances involved (up to 215 km) were too short for the recording of waves which had travelled below the Mohorovičić discontinuity. Data since obtained from artificial explosions continued to give definitely higher velocities in the so-called granitic layer than those found from earthquakes. Velocities of about 6.0 were found by Wood and Richter (1931) near Pasadena. Additional results are summarized in Table 1. The data indicate that in these regions the velocity of longitudinal waves in the "granitic layer" is about 6 km/sec, and that sometimes a second layer having velocities as high as 6½ to 7 km/sec exists in the uppermost layers of the earth; the general conclusion differs definitely from that drawn from earthquake waves. However, it must be considered that no low-velocity layer can be found from refracted explosion waves if it exists below a layer with higher velocity.

TABLE 1. VELOCITIES V OF LONGITUDINAL WAVES IN CONTINENTAL AREAS FROM ARTIFICIAL EXPLOSIONS
(d = depth range)

Region	Reference	d km	V km/sec	d km	V km/sec	d km below	V km/sec
Northern Germany	Wiechert (1926)	0–?	5.98				
Northern Germany	Brockamp (1931)	0–8	5.9	8–?	6.7		
Helgoland	Wilmore (1949) (a)	7–27	5.95			27	8.18
	(b)	6–14	5.57	14–30	6.5	30	8.18
Helgoland	Schulze and Förtsch (1950)	6–11	5.34	11–27	6.2–6.6	27	8.19
Black Forest	Reich et al. (1948)	0–21	5.9–6.0	21–31	6.55	31	8.2
Black Forest	Rothé et al. (1950)	2–20	5.97	20–30	6.54	30	8.15
Canadian Shield	J. H. Hodgson (1947)	0–17	6.15	17–36	6.45	36	8.20
New England	Leet (1936)	0–23	6.0			23	8
Near Washington, D. C.	Tuve (1948)	0–10	6.0–6.7	10–42	6.7–7.1	42	8.15
Southern California	Wood and Richter (1931)	0–?	6.0				
Near Corona, Calif	Gutenberg (1951a)	0–6	5.7–6.0	10–?	6½–7	40±	8.1–8.2
Near Corona, Calif.	Tuve and Tatel (1950)	3–14	6.3	14–?	6.8		

Until recently it was frequently assumed that either the numerical values of the velocities calculated from the earthquake observations or those from the artificial explosions were incorrect. In an attempt to combine both groups of findings, Schmerwitz (1938) suggested that the velocity in the earth's crust decreases linearly with depth from about 5.9 km/sec near

the surface to about 5.5 km/sec at a depth of about 50 km. Gutenberg (1943b) pointed out that in certain regions a layer of higher velocity may exist above a layer with somewhat lower velocity (5.6 km/sec) and suggested that the high-velocity layer may consist of sedimentary rocks with velocities higher than that in the underlying "granitic" material. Wood (1948) considered the possibility that increase of pressure with depth may increase the velocity in certain upper levels of the "granitic" layer, while below the increase in temperature more than offsets this effect.

The recent findings from blasts suggest a completely new interpretation of the "P-wave" in earthquake records. It no longer seems possible to combine the findings from blasts originating close to the surface with those from earthquakes on the assumption that thin crustal layers *of limited extent* and exhibiting wave velocities higher than that in the deeper "granitic layers" produce the differences between the crustal structure found from earthquakes and that deduced from surface explosions. It rather must be concluded that the velocity in the upper "granitic layer" is about 6 km/sec. It becomes difficult now to assume that in earthquake records $\bar{P}$, at least at the greater distances where it is not the first phase, is the direct wave through the "granitic layer". Its relatively low velocity suggests the possibility of a low-velocity layer at some depth below the surface of the "granitic layer". Since investigations concerning low-velocity layers in the atmosphere and in the ocean have shown the existence of the "sofar waves"[1] which produce considerable amplitudes at some distance from the low-velocity "channel", the possibility must be considered that starting at a certain distance, $\bar{P}$ is the effect of a wave of this type. If this assumption should be correct, records of artificial explosions at short distances would begin with waves which have travelled with a velocity of about 6 km/sec through the so-called "granitic layer"; at somewhat greater distances, say 100 km from the source, the first waves would have apparent velocities of 6 to $6\frac{3}{4}$ km/sec, which would be characteristic of the velocity of the material at a depth of roughly 10 km. In records of artificial explosions or of earthquakes originating near the surface, no phases could be expected which belong to refracted waves with their deepest point in the low-velocity layer; thus, no information on the velocity in this layer could be obtained from refracted waves, and no strong waves corresponding to $\bar{P}$ in the earthquake records could be expected in blast records at greater distances. The rather scanty explosion records taken at distances of 200 km or more seem to confirm this conclusion. Calculated depths in Table 1 of more than 15 km would require revision. Earthquake foci would be frequently in the low-velocity layer, below which, at an unknown depth, the wave velocity begins to increase again. Finally, at the Mohorovičić discontinuity, the velocity jumps to about 8.2 km/sec.

The next step is to find if such a picture would satisfy the travel times of P-waves in earthquakes. For this purpose travel-time curves have been calculated (Gutenberg 1951b) supposing that the velocity of longitudinal waves is 6 km/sec in the uppermost 5± km of the earth's crust, $6\frac{1}{2}$ km/sec below that down to a depth taken arbitrarily as about 12 km, and that there it decreases suddenly to 5.6 km/sec, which remains constant down to a depth of about 16 km, where the earthquake foci are assumed to be located. The earlier data on travel times of southern California arthquakes (Gutenberg, 1943b) can be madee to fit these calculated values with reasonable accuracy (Gutenberg, 1951b) if the origin times are on the average about $1\frac{1}{2}$ seconds later than was calculated in the earlier papers on the assumption that $\bar{P}$ is the direct longitudinal wave at distances up to at least 200 km. Such a change would not contradict any other observations, but it would remove the apparent early arrival of the transverse waves, which was explained, for example, in the case of the Long Beach earthquake, as an effect of the speed of faulting. It would follow that this speed does not exceed the velocity of transverse waves. Under the new assumption, the calculated travel-time curve for the direct P-wave would fit very well the observed travel times of

[1] From SOund Fixing And Ranging (Navy Department) using natural sound channels. See e.g. Ewing and Worzel (1948) especially figure 5, p. 19. The term "sofar-wave" is now generally used for waves travelling with little loss of energy along low-velocity layers or channels.

$\bar{P}$ in southern California earthquakes out to a distance of about 60 km, where under the previous assumption the travel-time curve of $\bar{P}$ is intersected by the travel-time curve of the wave with the symbol Py (apparent velocity slightly over 6 km/sec). It would seem now that the observed travel-time curve of Py is a part of that for the direct wave, which at very short distances is given by $\bar{P}$; amplitude observations for both waves are rather in favor of the new than of the old connection (Gutenberg, 1951b). Starting at the distance where the movement due to the energy radiating from the "sofar" channel (or some other cause) becomes observable, Py would precede this phase ($\bar{P}$ at a distance of roughly 50 km or more, depending on the conditions) by time intervals increasing with distance from a very small fraction of one second to many seconds. Thus, one supposed layer would be removed. No phase can now be considered the effect of a wave with its deepest point below the earthquake focus if it exhibits an apparent velocity smaller than that of about 6¼ km/sec found nearer the surface. Richter (1950) has investigated a number of records from smaller earthquakes in southern California during 1949, including data from four semi-permanent stations with highly sensitive Benioff-vertical seismographs which have been installed especially for such research. His results (no longer based on the assumption that at greater distances $\bar{P}$ is the direct longitudinal wave) are in excellent agreement with the results expected on the new hypothesis for the travel-times of the direct longitudinal wave. They give an "average" velocity of about 6.3 km/sec for this wave.

The velocities between the depth at which most shallow earthquakes originate (about 18 km) and the Mohorovičić discontinuity are difficult to calculate because of the low-velocity layer. P-waves which start at a rather large angle of incidence, which means not too far from horizontally, going upward or downward, would form a "sofar" channel with its characteristic behavior and may be the cause of $\bar{P}$ and similar waves at greater distances. Since no waves from artificial explosions could have their deepest point in this layer, additional details regarding it could be found only from recorded reflected waves, if one or both boundaries of the channel are formed by discontinuities. However, at least on the upper side of the channel, the velocity seems to change gradually with depth, so that no reflected waves would be produced.

In earthquakes, the travel times for the P_1 waves would be about 1½ seconds shorter on the new hypothesis than has been found previously. The same difference of 1½ seconds would have to be subtracted from all Pasadena travel-time curves of shallow shocks out to the greatest distances since the origin time was based on the findings for small distances. This result would agree very well with results from the Bikini explosion (Gutenberg and Richter, 1946), where such a difference was observed.

The calculated depth of the Mohorovičić discontinuity would probably be decreased by a few kilometers since the decrease in the travel times of P_n in earthquakes would be divided between the path going down to the discontinuity and the path of the wave coming up again at the end. On the other hand, changes in the velocities in the uppermost layers may partly balance this calculated change in depth. The new solution would continue to be in agreement with the observed reflections from the various layers. The new interpretation probably would remove the occasional result of a "negative" thickness of a layer calculated from earthquake data.

The suggested velocity distribution would have the effect that those surface shear waves which carry their maximum energy roughly within the uppermost 25 km could probably not strongly develop, since much of the energy of waves of the corresponding lengths would travel in a layer with the velocity decreasing downward. Such surface waves would have periods of about 6–10 seconds. Actually, surface waves with periods of less than about 12 seconds are observed very rarely in records of distant earthquakes.

The new results would also affect the calculated amplitudes of PP, pP, sP, etc. Waves with a period of 6 seconds in the crust (wave length of about 40 km) can be expected to be reflected with an angle of incidence depending appreciably on the wave velocity in the low-

velocity layer; however, for much shorter waves the angle of incidence at the surface is determined principally by the higher velocity in the upper layer. Again, this corresponds well with the facts. Gutenberg and Richter stated, for example, (1935, p. 311) "that PP/P appears too small on short-period instruments" (which record mainly short-period waves).

Such a low-velocity layer, in many areas of the continental crust at least, may be explained either by assuming a succession of relatively thin layers of different material in many regions of the earth's crust, or by the effect of physical processes, or by both. In a given material, the velocity of the elastic waves is mainly an effect of pressure and temperature (Anderson, 1938). Experiments indicate that the pressure tends to increase the wave velocity in a given material appreciably in the upper few kilometers, but much less below about 10 km. Few data are available for the effect of temperature, mainly for quartz, and to a lesser degree for diabase. They indicate a decrease in wave velocity with increasing temperature, followed by an increase. Young's modulus for quartz as a function of temperature under normal pressure after data compiled by Birch *et al.* (1942, p. 70) and based on the research of a number of investigators, is shown in Figure 1.

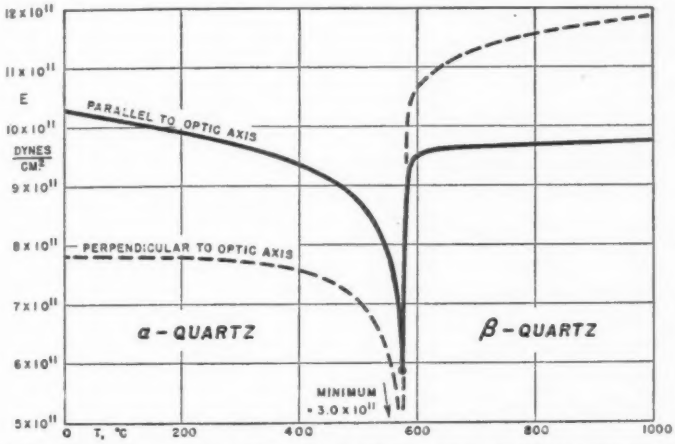


FIGURE 1.—YOUNG'S MODULUS E AS A FUNCTION OF TEMPERATURE AT NORMAL PRESSURE IN QUARTZ
After data given by Birch (Birch *et al.*, 1942).

The effect of pressure on the temperature of transition from alpha quartz to beta quartz was investigated by Gibson (1928) and Yoder (1950). Yoder's results are given in Figure 2, together with the temperature as a function of depth according to Urry (1949). The two curves intersect each other at a point corresponding to a depth of roughly 25 km. Von Wolff (1930), on the basis of older data, estimated the depth of the transition as about 20 km. Daly (1930) has suggested that the transition from alpha to the beta quartz is the reason for one of the discontinuities in the earth's crust, giving a higher velocity at a depth below about 30 km than above that level. The curve in Figure 2, giving the temperature as a function of depth has been found differently by others. Their results would lead to a greater depth for the transition. However, this depth is of no importance in the present paper.

Unfortunately, no data are available concerning the effect of pressure on Young's modulus for the two varieties of quartz. Supposing that the effect is similar to that in rocks and assuming an average change in temperature with depth (estimates of various authors do not differ very much in the range involved), the following changes in the velocity of longitudinal waves with depth in *pure* quartz would result:

Between the surface and a depth of 10 km, the velocity should increase because of the prevailing effect of increasing pressure and hence increasing moduli of elasticity. Between the depths of 10 and 15 km, a maximum velocity would be reached, and the velocity should decrease below, owing to the decrease of the elastic coefficients in approaching the transition from alpha to beta quartz (Fig. 1). A minimum velocity, noticeably lower than that near the surface, can be expected at the transition point, at a depth of 25 km or more, and below, within a depth range of 1 or 2 km, the velocity should increase to a value roughly 10 per cent higher than the shallower maximum. A slow increase in velocity with depth should follow. The general trend corresponds well with our hypothesis, though it is probable that the material at the depth of the transition from alpha to beta quartz is different from that at the depth where the decrease in velocity begins.[2]

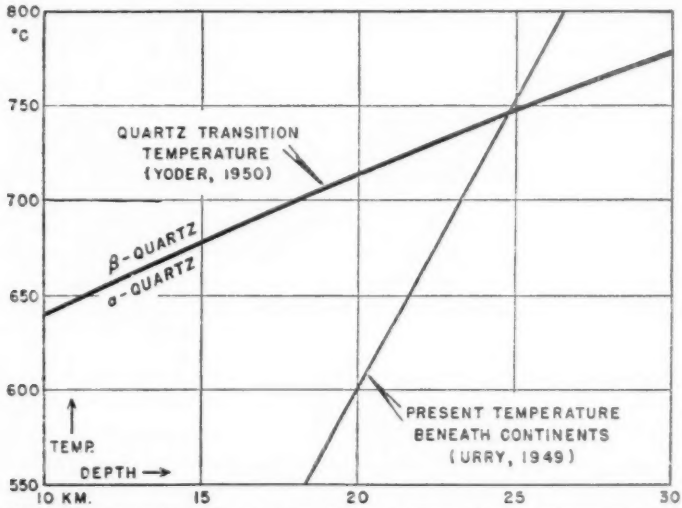


FIGURE 2.—TEMPERATURE OF TRANSITION FROM ALPHA TO BETA QUARTZ (AFTER YODER, 1950) AND PRESENT TEMPERATURE BENEATH THE CONTINENTS (AFTER URRY, 1949) AS A FUNCTION OF DEPTH

The accumulation of earthquake foci in the low-velocity layer may find an explanation in a relatively small breaking strength there. Unfortunately, few similar data are available for other constituents of rocks. Birch and Dow (1936) have described laboratory determinations of the elastic behavior of diorite under pressures up to 10,000 kg/cm^2 combined with temperatures of 0 to over 400°C. Using the resulting values of the compressibility, they calculated on reasonable assumptions regarding temperature and pressure in the earth's crust that the velocity of longitudinal waves in pure diabase would increase by almost 1 km/sec in the upper 6 km, decrease slightly below, and resume a slight increase with depth approaching a depth of 15 km. At 40 km, the velocity

[2] Note added in proof: In a recent paper Yoder (Am. Geophys. Union, Tr., vol. 31, 1950, p. 834) points out that probably "near the depths at which the transition might occur the rocks now thought to exist there are either very low in quartz or quartz free" and considers this the chief objection to the present hypothesis. However, Yoder overlooks here the fact (pointed out by him on page 833) that it is not the transition point but that at which the decrease in elastic constants begins which is of importance in the present hypothesis. For quartz, this critical point can be expected at a depth of 10 to 15 km, where the temperatures assumed by various authors differ only insignificantly. The minimum decrease in wave velocity v which is required for formation of a sofar-channel (in which no seismic waves arriving at the surface have their deepest point) is very small. It is given by $dv/dr = v/r$ which in our case leads to a decrease in velocity with depth by a minimum amount of only about 0.02 per cent per km. The corresponding required minimum decrease in the elastic constants is only about 0.04 per cent per km.

would be about ½ km/sec higher than at 15 km. Similar experimental data for feldspar and other constituents of the material in the earth's crust are necessary to establish the details of the low-velocity layer at a depth below about 15 km.

The fact that the low-velocity layer is not attributed to a world-wide relatively thin layer of a different material, but is considered to result from a natural physical property, is in favor of the hypothesis.

The new results show quite clearly that our knowledge of the earth's crust is much less than was assumed about 10 years ago. Especially, the increase in velocity (Table 1) from about 6 km/sec below the sediments to 6½ or 6¾ km/sec at a depth of about 10 km seems to be definitely beyond the amount to be expected in granitic material as an effect of pressure (even neglecting the effect of increasing temperature). The frequently assumed average thickness of the so-called "granitic layer" of 20 km or more was based mainly on the apparent velocity of 5.6± km/sec of P-waves in records of earthquakes distant up to about 500 km from the station. The new results indicate that it is not permissible to calculate the thickness of the "granitic layer" by use of travel times of the P-wave in earthquake records. In many areas covered by experiments, the layer with a velocity of longitudinal waves near 6 km/sec—which is to be expected in "granitic" material (Birch *et al.*, 1942, p. 80, 95) but does not prove its existence—does not extend below a depth of 10 km. The only regions in Table 1 where this layer may extend to greater depth are the Black Forest and the Canadian Shield.

For the layer on top of the Mohorovičić discontinuity, usually material of the type of gabbro or diabase is assumed, for which experiments (Birch *et al.*, 1942, p. 80) give velocities near 7 km/sec for longitudinal waves. For the material below the Mohorovičić discontinuity, artificial explosions (Table 1, last column) as well as recent calculations from P_n in earthquakes indicate a velocity of 8.1 to 8.2 km/sec, which corresponds to similar velocities observed in dunite. In continental areas, the depth of the Mohorovičić discontinuity varies between 10 or 15 km in some coastal areas to possibly as much as 60 km under high mountain ranges (Gutenberg, 1943a).

There is no doubt that the structure of the ocean bottoms differs materially from that in the continents. But, in addition, there are marked differences among the various oceans. E. Suess (1888) emphasized the difference between Atlantic and Pacific structure from geological evidence. Petrological differences resulted in establishing the Marshall (andesite) line which separates the inner Pacific Basin with prevailing olivine-rich alkali basalt from the outer area where pyroxene-andesite is prevalent. Where the andesite line could be established, it coincides with the inner boundary of the earthquake belt surrounding the Pacific Basin. There are possibly outlying areas with Pacific structure, such as the Caribbean and the area inside the southern Antillean arc and perhaps an area in the Arctic. On the other hand, certain parts of the Pacific Ocean near its borders, for example in the region between South America and the Easter Island rise, or between the Mariannas and the Asiatic continent, show indications of less basic layers (*See*, Gutenberg and Richter, 1949).

The areas characterized by andesitic material exhibit noticeable differences in structure. Since apparent differences in opinion were frequently caused by different use of the terms "sial" and "sima", the writer, following a suggestion made by L. H. Adams, has changed his use of these terms (previously sial to include granite and basalt, sima for ultrabasic material) to comply with the following definitions: "Simatic rocks" refers to basalt and gabbro (also to diabase and dolerite); rocks less basic than these are called "sialic rocks"; rocks more basic than simatic rocks are called "ultra-simatic".

While there is no proven correlation between the rock type and the velocity of longitudinal waves under conditions near the surface, the following ranges are occasionally used as a guide: sialic rock up to about 6¼ km/sec; simatic rocks over 6¼ but less than 8 km/sec; ultra-simatic rocks 8–8¼ km/sec. The Mohorovičić discontinuity is then the upper boundary of the ultra-simatic rocks.

Data on velocities of body waves in the crustal layers beneath the oceans are scanty.

Results by Angenheister (1921) indicating relatively high velocity in the bottom of the Pacific basin near Samoa are not reliable; later determinations of epicenters and origin times have shown that some of the results which he found are inaccurate by an amount of a multiple of the deviation. In other instances, the shocks which he considered all originated at shallow depth had an appreciably larger depth of focus. Raitt (1949) observed a "basement" velocity of about 6½/ km/sec beyond the continental slope off the southern California coast and a velocity of about 8 km/sec starting at a depth of about 5 km. Such a type of structure can be expected for the Pacific Basin considering the volcanic islands and the bottom topography (sea mounts). In this case, the Mohorovičić discontinuity would be at a depth of a few kilometers with a combination of sediments, erupted and perhaps other simatic material above it.

The Atlantic may be divided into five distinct units which are from west to east: A submerged coastal plain, the western basin, the Atlantic ridge, the eastern basin, and an eastern submerged coastal area. In the North American coastal plain, velocity measurements by Ewing *et al.* (1950a) indicate velocities of the order of 3 to 4 km/sec in semi-consolidated sediments and of 5 to 6 km/sec in the basement. In the western basin near Bermuda, Ewing *et al.* (1950b) found a velocity of 7.58 km/sec below a sedimentary layer about 1½ km thick.[3] The "granitic" and "intermediate" layers were absent. This velocity is so much less than the 8.1–8.2 km/sec in the ultrabasic material below the Mohorovičić discontinuity in the continents (Table 1) that it indicates different (probably less basic) material. Ewing (1950) believes it would probably be olivine basalt. No information on a velocity of over 8 km/sec (to be expected below the Mohorovičić discontinuity) exists from records of explosions in the Atlantic,[3] and no information of any kind on velocities in the Atlantic ridge. In the eastern basin, about 600 km west of Ireland, velocities of 5¼ km/sec below a depth of about 2½ km and of 6½ km/sec below 5 km were reported by Hill and Swallow (1950). There seems to be no information on the depth of the Mohorovičić discontinuity in this area.

Summarizing the data, there is some indication that, in an area in the eastern part of the Pacific Basin, the ultrabasic material with a velocity of about 8 km/sec begins at a depth of only a few kilometers, but that in an area in the western Atlantic basin material with a velocity of 7½ km/sec, definitely less than that of 8.1–8.2 km/sec below the Mohorovičić discontinuity in the continents, seems to exist at the corresponding depth. In the eastern Atlantic Basin the velocities may even indicate some sialic crustal material. The depth at which the velocity begins to reach or exceed 8 km/sec has not yet been established by use of artificial explosions in any part of the Atlantic. Earthquake records indicate that at a depth of roughly 60 km a velocity of about 8 km/sec exists everywhere.

While, except for the topography, there are no clear features outlining the boundary between different structural types in the Atlantic or near its coasts, there are clear geophysical phenomena which accompany the discontinuity between the Pacific Basin and the surrounding continental area, consisting of several groups at various distances (Gutenberg and Richter, 1949). Nearest to the boundary are great ocean deeps, negative gravity anomalies, and shallow earthquakes. Next follow a narrow belt of earthquakes at depths of about 70 to 150 km and lines of active volcanoes. Still farther from the boundary are shocks at increasing depths down to 700 km. No other region on earth now shows this complete sequence of phenomena, nor are there earthquakes deeper than 300 km. except along this belt. None of these phenomena occur along the coastal areas of the Atlantic and Indian Oceans, nor is there any other known indication of a discontinuity between the bottom there and the surrounding continents.

The velocity of surface waves across the various units of the earth's crust have been used repeatedly for the investigation of struc-

[3] Note added in proof: Additional data and recalculations give a velocity of about 7.2 km/sec instead of 7.58 km/sec according to Dr. Ewing (oral communication, January 1951). He also finds indications that this layer is roughly 5 km thick and that a velocity of 8 to 8½ km/sec prevails below it. (No change was made in the text, especially not in the following paragraph.)

ture. In a medium which is not homogeneous, velocity of surface waves depends on the period. Short waves are propagated only in a thin layer, the properties of which determine their velocity, whereas the energy of long waves is propagated in a relatively thick layer. In each case the disturbance extends from the surface to a depth that is a small multiple of the wave length and then continues to fall off exponentially with depth. This fact can be applied to both Love waves and Rayleigh waves. However, study of the latter is more complicated, both from the theoretical point of view and because of the difficulty in identifying them properly, since they have a smaller speed than the Love waves and therefore are usually superposed by Love waves. Although the findings of various authors differ (Fig. 3), the indication—though not very strong—is that the velocity of short Love-waves is highest in the bottom of the Pacific Ocean, slightly smaller across the bottom of the Atlantic and Indian Oceans, and appreciably smaller across the continents. Calculations of the thickness of the layers and the respective wave velocities are very complicated, but the curves indicate that, in the uppermost part of the continental crust, the layers with relatively low velocity are thickest, that they are much thinner in the Atlantic and Indian oceans, and still slightly thinner in the Pacific. However, the difference between the surface wave velocities across the Atlantic and those in the Pacific could be a consequence of the relatively wide submerged coastal areas and the mid-Atlantic ridge.

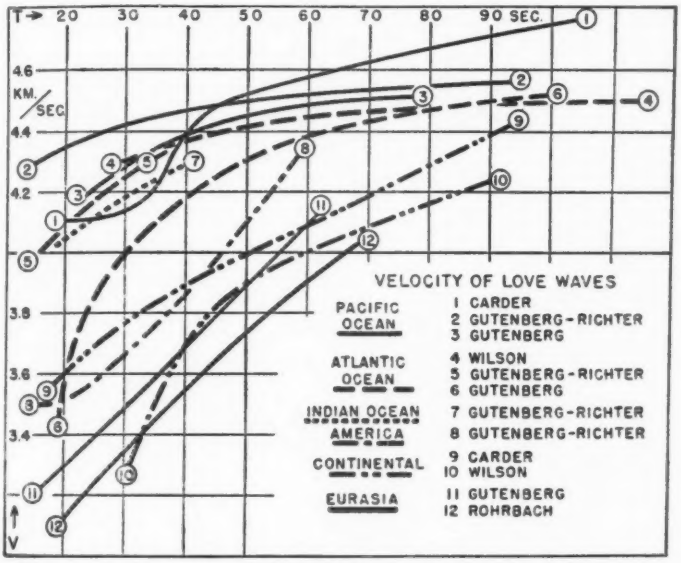


FIGURE 3.—OBSERVED GROUP VELOCITY OF LOVE WAVES AS A FUNCTION OF PERIOD IN VARIOUS UNITS OF THE EARTH'S CRUST

After Carder (1934), Gutenberg (1925), Gutenberg and Richter (1936), Rohrbach (1932), and Wilson (1940).

Significant results have been found through the study of the change in amplitudes of surface waves with distance. Surface waves travelling either under the Pacific only, or through the continents only, show relatively little loss of energy. Surface waves crossing the Atlantic and parts of continents on both sides again show about normal absorption. However surface waves passing for longer distances along the boundary of the Pacific basin, as it is determined by the andesite line or the earthquake foci, show a great loss of energy amounting in extreme cases to about 90 per cent

(Gutenberg, 1945). In this way it is possible to locate approximately the boundary of the Pacific structure. Observations do not indicate any discontinuities of similar magnitude anywhere along or in the Atlantic and Indian oceans.

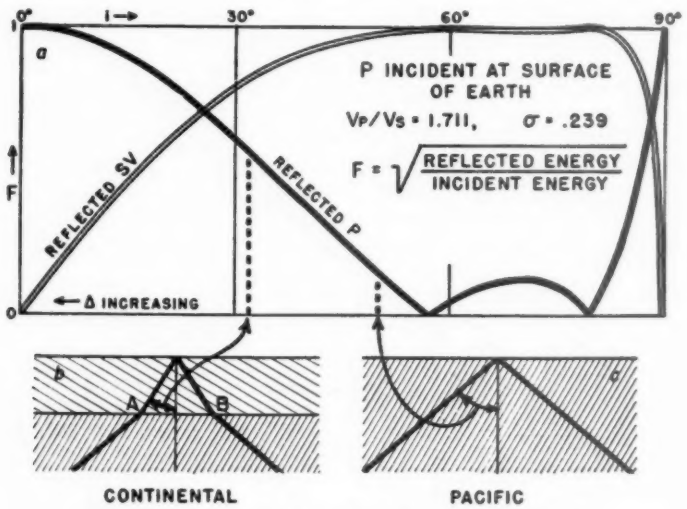
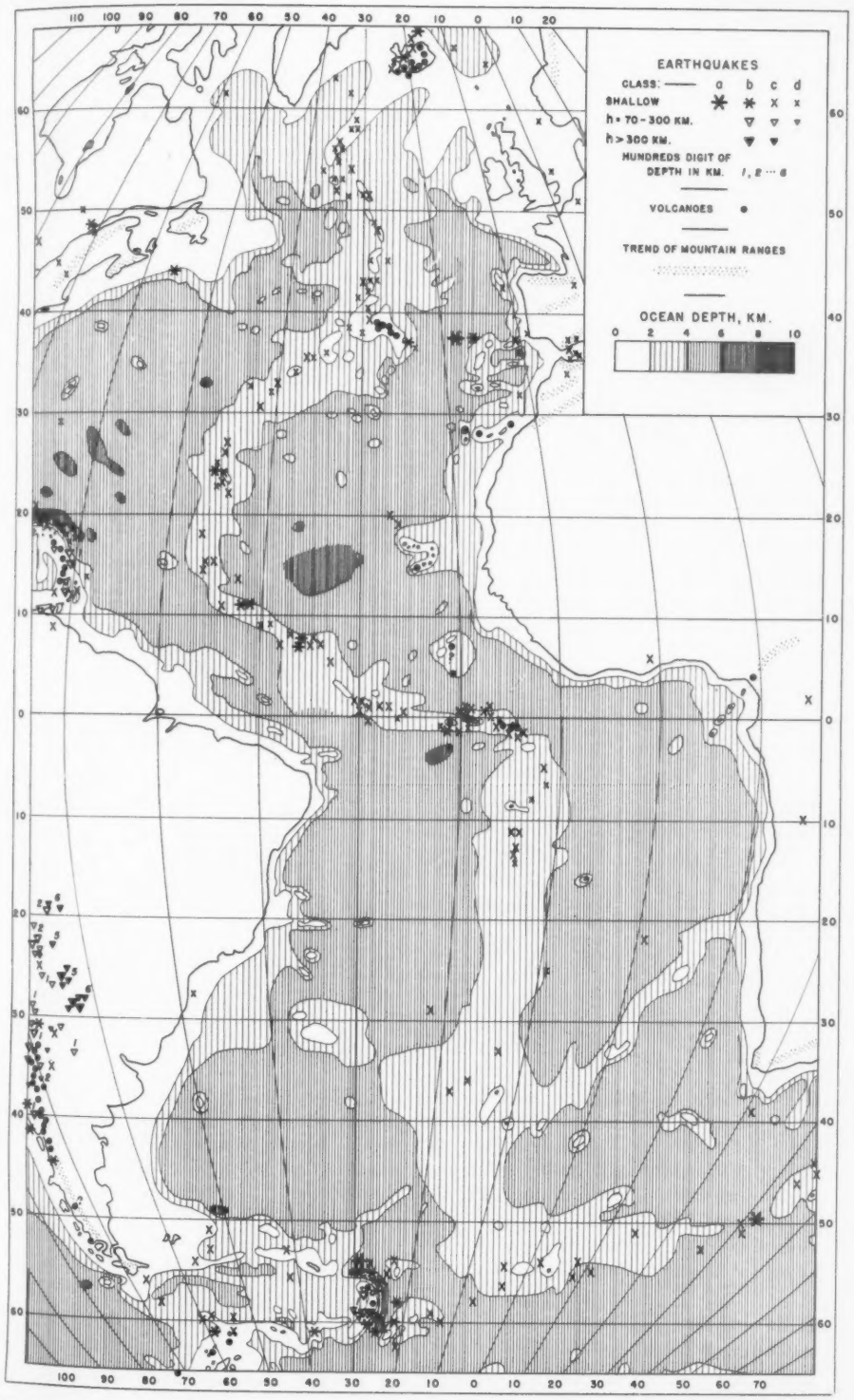


FIGURE 4.—SKETCH OF PATHS OF REFLECTED LONGITUDINAL WAVES

Lower right: Pacific reflection; lower left: continental reflection; top: ratio of reflected to incident amplitudes as a function of angle of incidence (i) and epicentral distance (△). At the Pacific bottom, reflection of a PP-wave arriving at a certain distance occurs at a greater angle i (in the figure about 50°) than at the continental surface (i about 32°) due to lower velocity in the latter. The upper part shows that, in the first case, the amplitude of the reflected wave is about 55 per cent (dotted line) of the incident wave, in the second only about 15 per cent. The actual values depend on the velocity near the reflecting surface and Poisson's ratio. The figure indicates that, for a Poisson's ratio of 0.239, little energy is reflected into PP for waves arriving under angles of incidence between about 50° and 80°, and that in such cases practically all the energy goes into the reflected transverse wave (PS).

Longitudinal waves reflected under the Pacific basin, on the average, have relatively smaller amplitudes than the corresponding waves reflected under the continents and under the other oceanic areas. This may be at least partly due to the fact that in the bottom of the Pacific Ocean the wave velocity is higher, and consequently a smaller percentage of the energy is reflected there (Fig. 4). There may also be a difference in Poisson's ratio. Many seismograms from stations in all parts of the world have been examined for studies of amplitude ratios. Continental reflections for example have been found in the southeastern Pacific at about 25° south, 85° west; 20° south, 90° west; and 9° south, 91° west (Gutenberg and Richter, 1935, p. 315; 1939, p. 316). All reflections from the ocean floor north- north-east of the Galapagos Islands are "pacific". In exceptional cases, it is possible to locate the boundary between those structures by this method. It can not be used, however, to determine the thickness of the layers with lower velocity. This thickness must be at least a large fraction of the wave length of the longitudinal seismic waves, which usually is between 20 and 30 km. The method gives information as to structure only near the point of reflection.

The importance of seismic phenomena for the determination of the boundary of the Pacific has been mentioned. It is noteworthy that the earthquake belt marking the boundary of the Pacific (Fig. 5) contains about 80 per cent of all shallow earthquakes, an even larger fraction


EARTHQUAKES
CLASS: a b c d
SHALLOW
h = 70 - 300 KM.
h > 300 KM.
HUNDREDS DIGIT OF DEPTH IN KM. 1, 2 ··· 6
VOLCANOES
TREND OF MOUNTAIN RANGES
OCEAN DEPTH, KM.
0 2 4 6 8 10
110 100 90 80 70 60 50 40 30 20 10 0 10 20
100 90 80 70 60 50 40 30 20 10 0 10 20 30 40 50 60 70

EPICENTERS OF EARTHQUAKES AND OTHER GEOPHYSICAL PHENOMENA IN THE ATLANTIC AREA

After Gutenberg and Richter (1949)

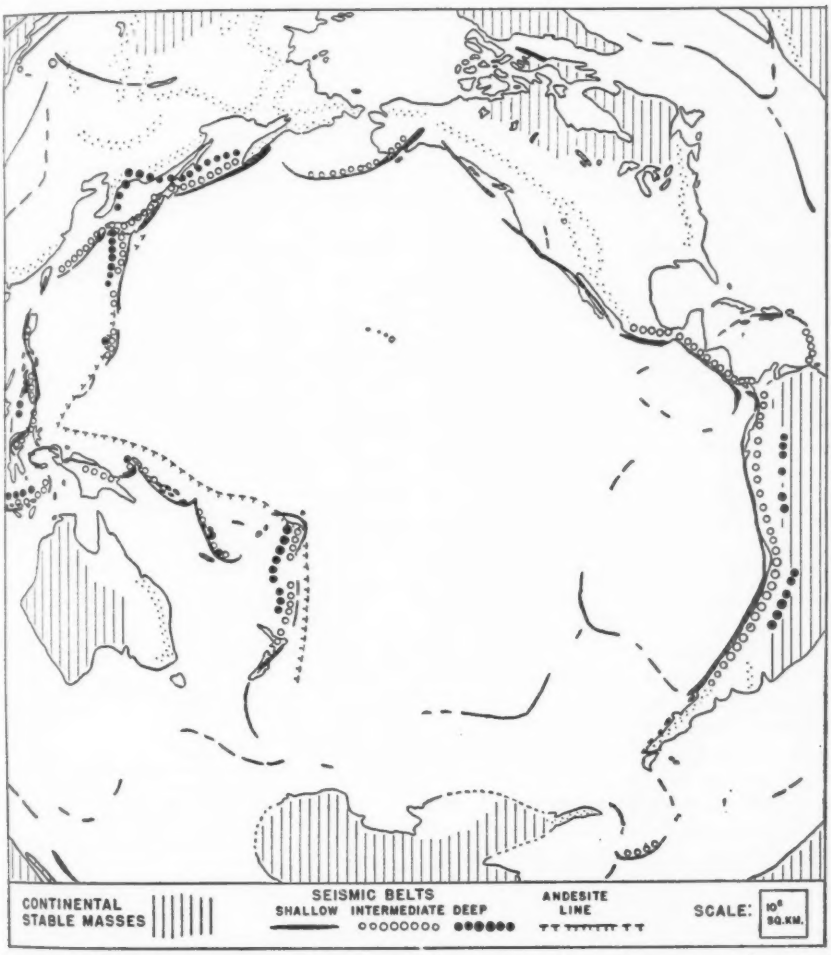


FIGURE 5.—EARTHQUAKE BELTS SURROUNDING THE PACIFIC AND OTHER GEOPHYSICAL PHENOMENA
After Gutenberg and Richter (1949).

of the shocks between 60 and 300 km, and all shocks without exception of focal depth greater than 300 km. No such earthquake belt surrounds the whole or parts of the Atlantic (Pl. 1) or Indian Ocean. In both, active belts of shallow earthquakes, including only rarely major earthquakes, follow ridges inside these oceans, similar to the conditions in the continents where earthquake epicenters follow mountain ranges. Contrasting with the continents and the Indian and Atlantic oceans, practically no earthquakes are known to have occurred inside the Pacific basin proper, with the exception of the Hawaiian area (Gutenberg and Richter, 1949). The belts of earthquakes and other phenomena outline primarily zones along which certain processes are active. But they also define stable blocks—the largest being the Pacific basin—and their present boundaries.

The combination of all available results leads to the following tentative conclusion

regarding the present structure of the earth below the sediments:

In the continents, the crust is formed by a layer of sialic rock of a thickness varying from place to place, with an average of the order of 10 km. Below it, down to the Mohorovičić discontinuity at a depth usually between 30 and 40 km, but deeper under at least some mountain areas, are simatic rocks. Below the Mohorovičić discontinuity is ultra-simatic material. The thickness of the layer of sialic rocks decreases under the shelves of the Atlantic and Indian Oceans which except for sediments have no continuous sialic layer under their deep basins, but probably a layer of simatic material at least in a large fraction of their area, and ultra-simatic rocks below it. The depth where this is reached (Mohorovičić discontinuity) is noticeably smaller than under the continents. A similar structure may be expected under parts of the Pacific Ocean outside the actual Pacific basin. Under the Pacific basin (outlined by the Marshall line, and approximately by the inner earthquake belt, where there is no petrographic evidence) ultra-simatic material is in general noticeably less than 10 km (about 5± km?) below the surface; under most parts of the Pacific basin there is no continuous layer of sialic material, but a combination of sediments and erupted simatic material forms the upper layers.

For the present major tectonic and geophysical processes, the boundary of the Pacific basin (Marshall line) seems to be more important than any other crustal boundary and to be frequently, if not everywhere, the surface trace of a sharp, deep-reaching discontinuity (Benioff, 1949); in contrast, the transition from areas of continental type to those of "Atlantic" type seems to be usually rather gradual.

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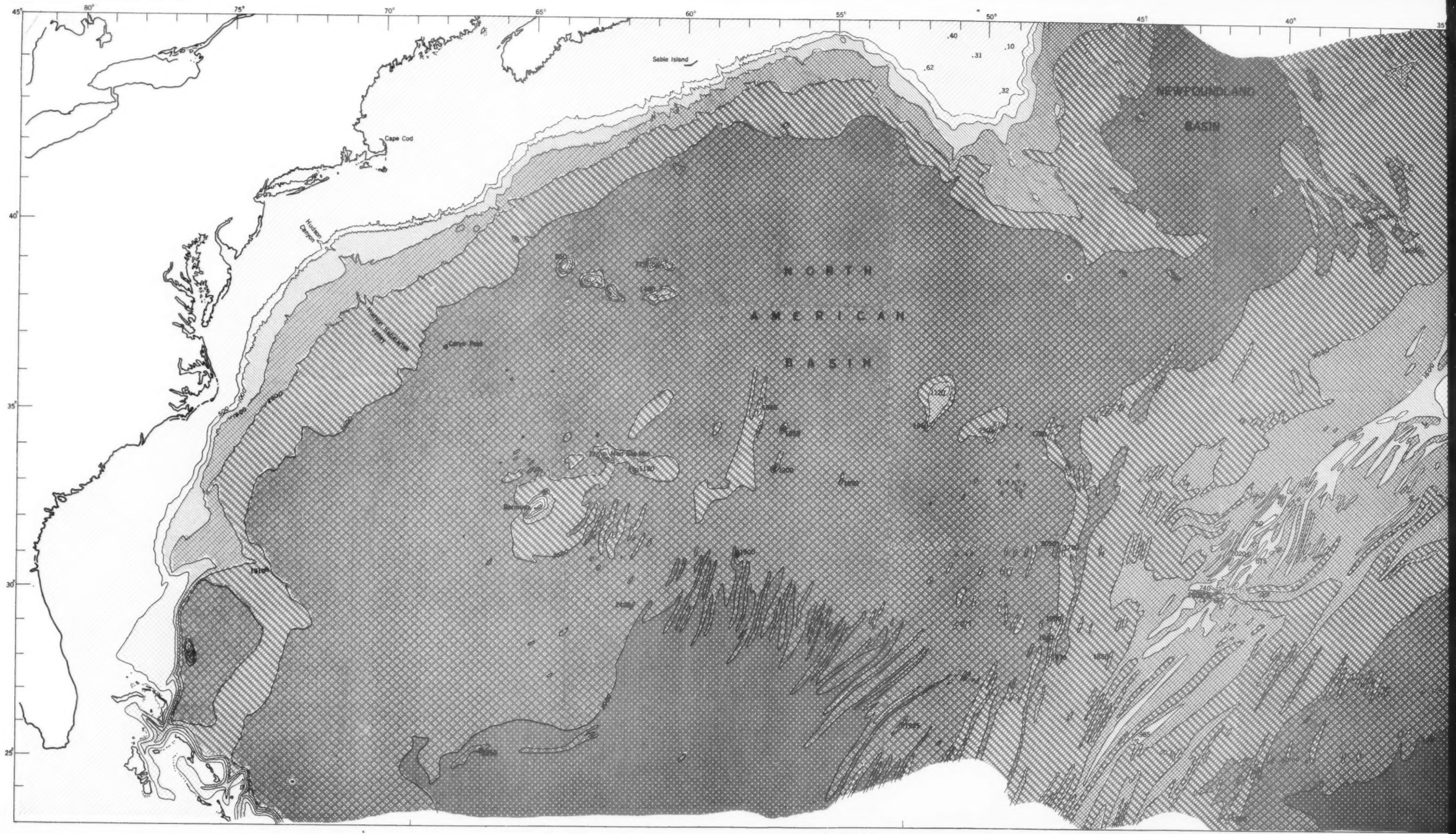
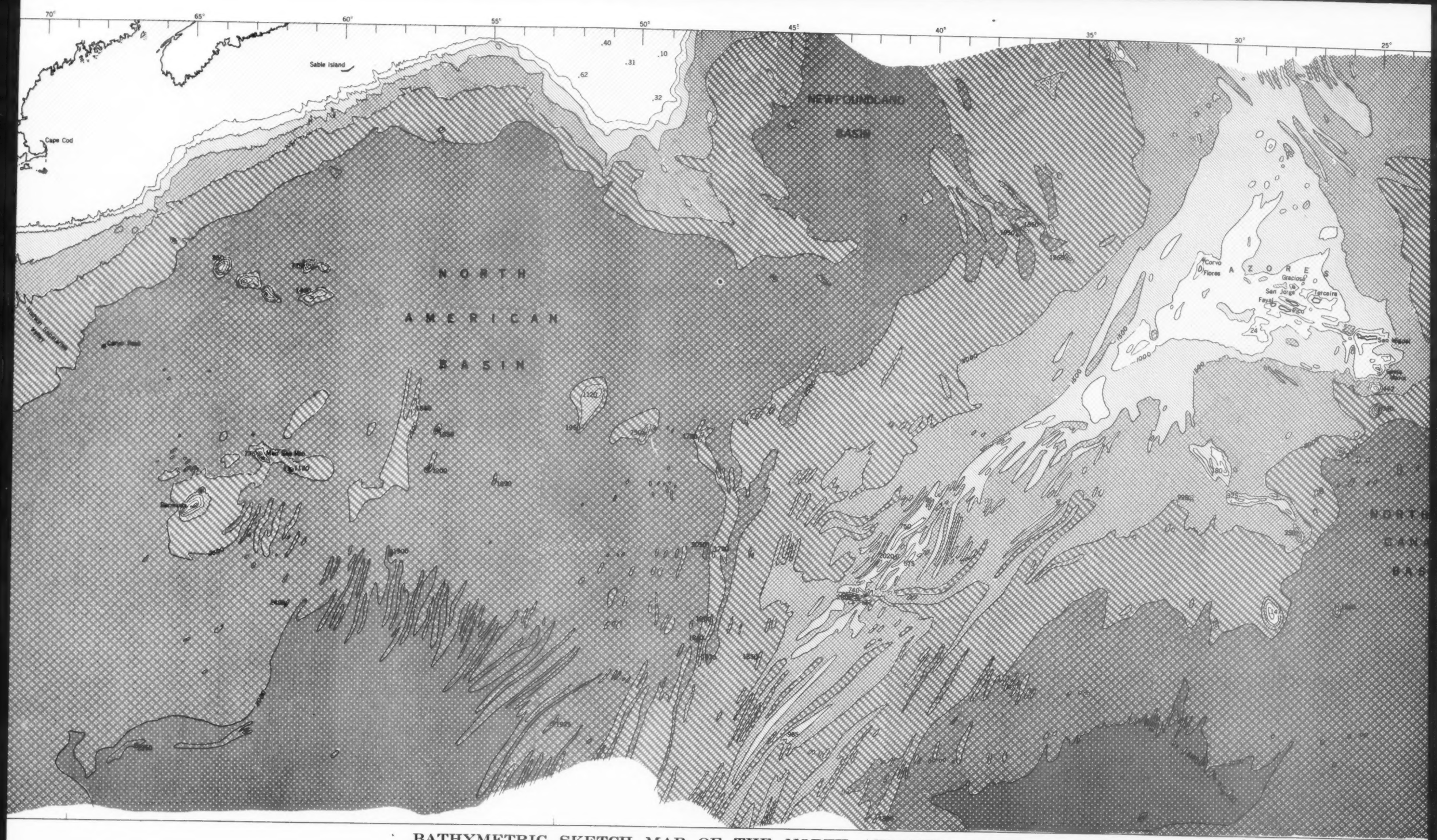


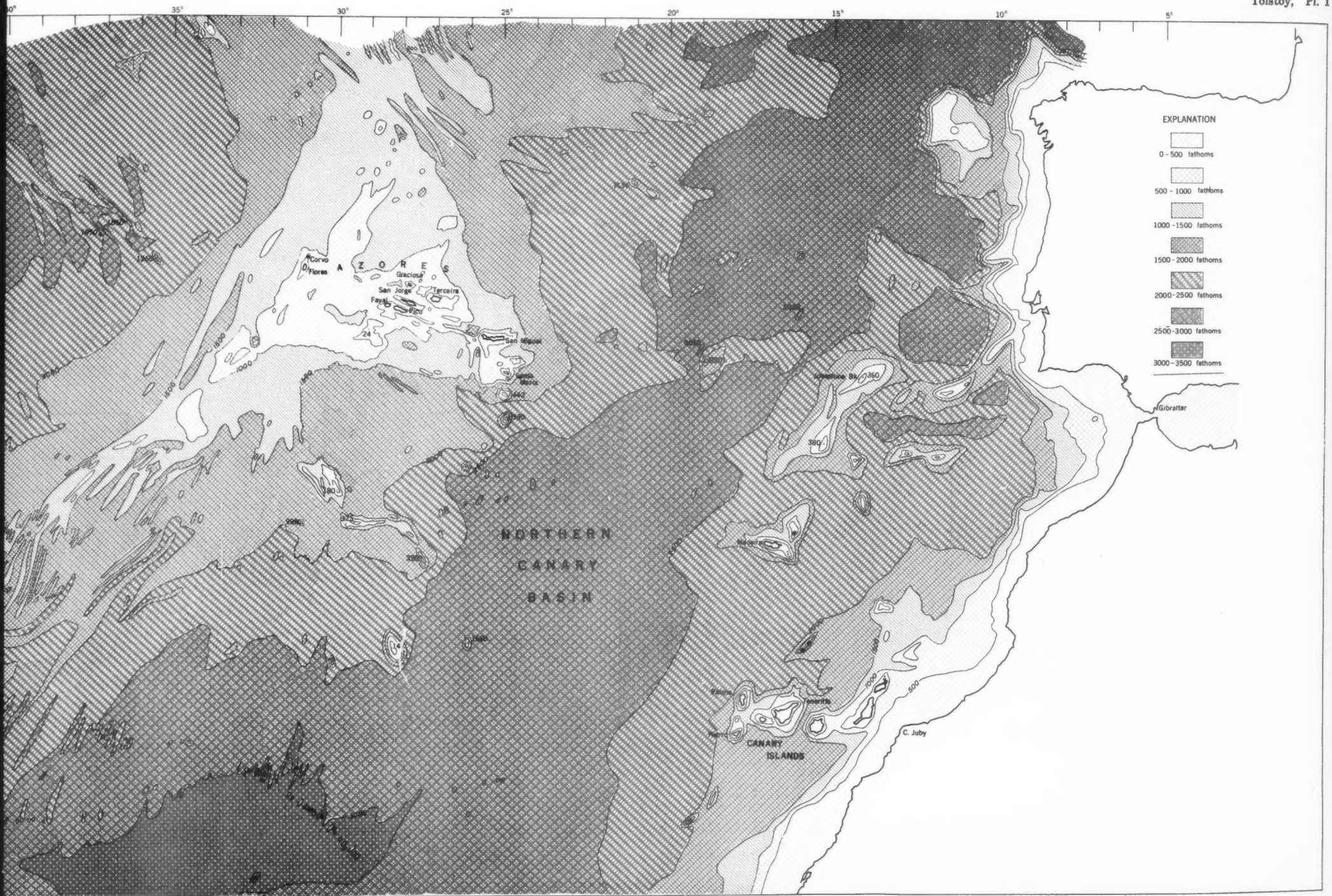
BATHYMETRIC SKETCH MAP OF THE NORTH ATLANTIC OCEAN, BETW

Contour interval 500 fathoms. The sounding lines used are shown in Plate



BATHYMETRIC SKETCH MAP OF THE NORTH ATLANTIC OCEAN, BETWEEN LAT. 45° N. AND 23° N.

Contour interval 500 fathoms. The sounding lines used are shown in Plate 2



NTIC OCEAN, BETWEEN LAT. 45° N. AND 23° N.

ding lines used are shown in Plate 2

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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SUBMARINE TOPOGRAPHY IN THE NORTH ATLANTIC

By Ivan Tolstoy

Abstract

In the North Atlantic one may distinguish several types of topography, each of which is characteristic of certain portions of the ocean floor.

The continental slopes are now known to show in many cases a steplike succession of horizontal or imperceptibly sloping shelves or terraces. They are dissected by systems of submarine canyons, the pattern of which is such as to suggest to many a subaerial origin. A recent investigation of the Hudson submarine valley by the Atlantis has shown that this valley extends at least as far as the 2500 fathom curve, 300 miles out to sea, and shows features suggesting a subaerial origin.

The broad basins are characterized by smooth floors or plains covering areas of more than 200,000 square miles out of which there rise large sea mounts, isolated or in groups. Thus plains at a depth of 2900 fathoms occupy the floors of the North American and the North Canary basins.

The floor of the northwestern part of the North American basin, north of Bermuda, is characterized by a plain at 2650 fathoms. Sea mounts of various sizes and shapes rise out of these otherwise smooth plains. There is a conspicuous group of flat-topped sea mounts rising from the 2650-fathom plain of the western North American basin. They rise approximately to 800 fathoms and string out toward the southeast, roughly from the direction of Cape Cod. They show evidence of terracing on their flanks.

The central part of the Atlantic Ocean is characterized by a long submarine mountain range, extending from Iceland to a point southwest of the Cape of Good Hope, known as the Mid-Atlantic Ridge. Numerous fathograms across the Ridge have shown that it is characterized by a high central zone, or Main Range, consisting of parallel ridges following the general trend of the Ridge and rising in many places to less than 800 fathoms below the surface of the ocean.

Between 1600 and 2500 fathoms the flanks of the Ridge are characterized by a succession of flats which for the lack of any better term have been called terraces. Reflection-shooting studies by Ewing and Press and others have shown that these terraces are or have been areas of greater deposition.

Finally, between this Terraced Zone and the 2900-fathom plain, one may usually recognize a mountainous area standing out as a distinct physiographic province.

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Introduction

Until the relatively recent development of echo-sounders that could record to depths greater than 2000 fathoms, there were few reliable data on the topography of the North Atlantic Ocean bottom. A few regions adjacent to land such as the continental shelf and the continental slope of the northeastern United States (Veatch and Smith, 1939) and the Azores bulge (Wüst, 1939) had been worked out sufficiently to give us an understandable morphological and structural picture of some small sections of the North Atlantic sea floor. Scientists and mariners had also been aware of some of the more outstanding features such as the Mid-Atlantic Ridge, the huge banks off the west coasts of Spain, Portugal, and North West Africa, and some of the great gashes cutting the continental slopes on both sides of the Atlantic such as the Hudson, Congo River, and Nazare canyons (Hull, 1912; Spencer, 1903). The vagueness of most of the sounding data and the lack of complementary geophysical information (such as composition and thickness of the sedimentary cover, elastic-wave velocities in the suboceanic crust, rock samples, etc.) have resulted in numerous and sometimes wondrous hypotheses concerning the history of the Atlantic Ocean. Though we are still far from understanding this fascinating but highly intricate problem, the work now being done by Columbia University and the Woods Hole Oceanographic Institution is beginning to extend into the oceans factual data heretofore limited to the continental shelves of the United States. The same situation had prevailed for many years in the Pacific Ocean, until the outstanding work of H. H. Hess, F. P. Shepard, the Scripps Institution of Oceanography, H. W. Murray, and many others began to clarify the morphology of the Pacific Ocean floor.

This paper illustrates the existence of oceanic features some of which were unknown and others only suspected. The author hopes it will contribute to our understanding of the history of the Atlantic Ocean, if only by emphasiziny the gaps in our knowledge and the complexity of the problem.

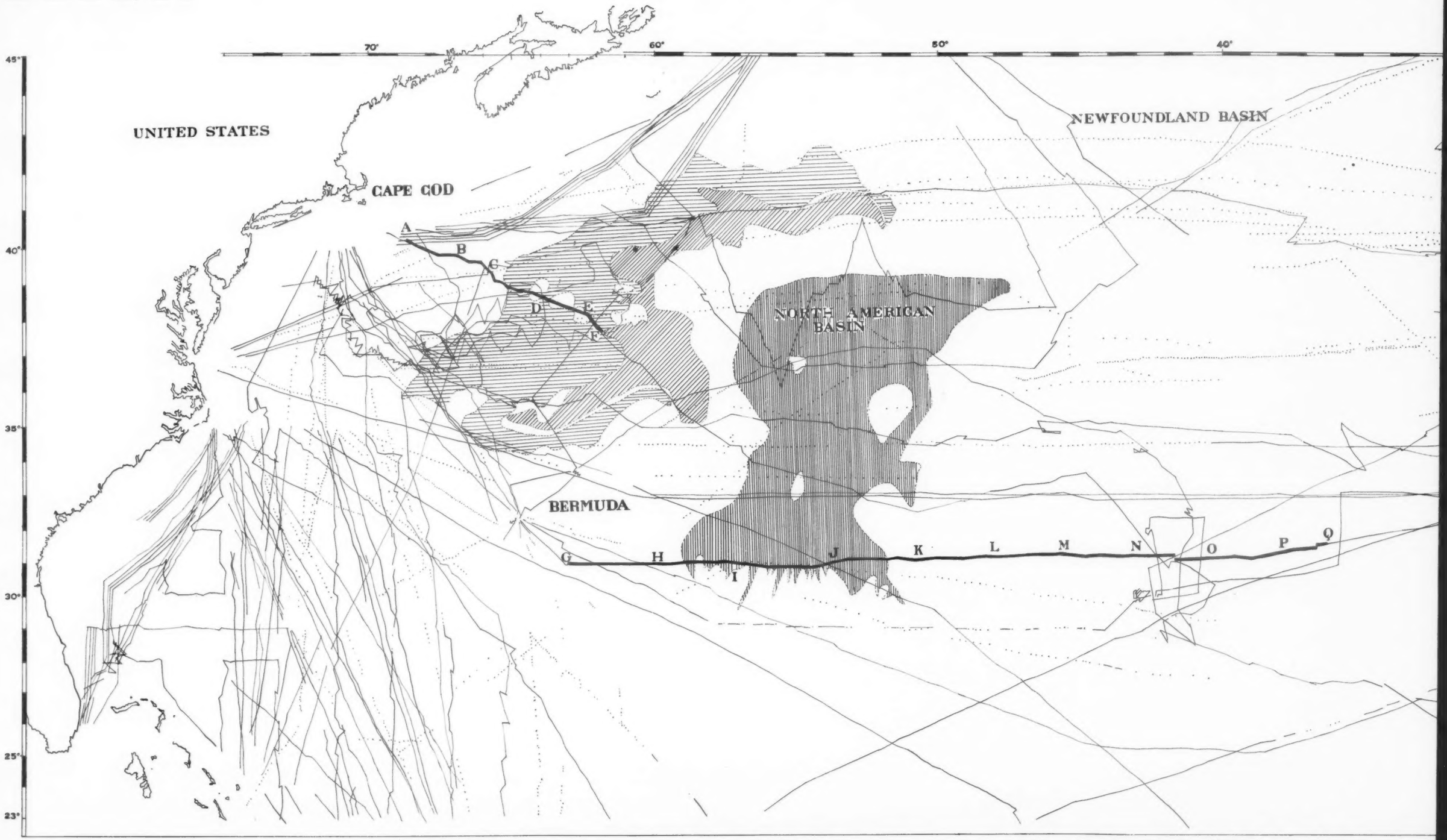
Sources and Acknowledgments

Some of the first continuous fathograms in water deeper than 2000 fathoms were obtained in 1946 by the R/V Atlantis of the Woods Hole Oceanographic Institution. Since that year over 100,000 miles of ocean bottom have been reproduced on fathograms, by the Atlantis and by a few other vessels. Numerous closely spaced discrete soundings have also been obtained by vessels doing work for the U. S. Navy Hydrographic Office.

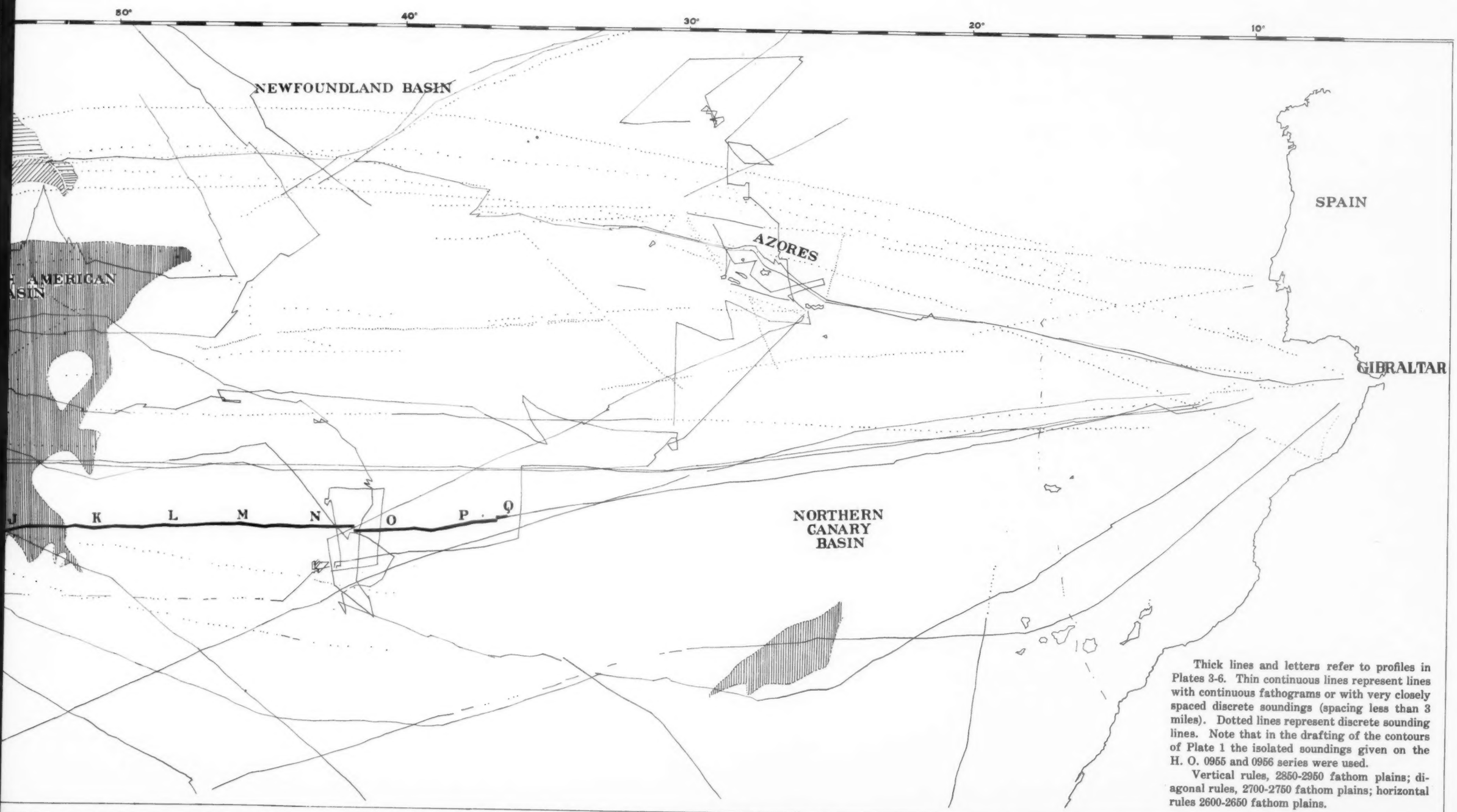
The author is grateful to the Woods Hole Oceanographic Institution and its personnel who made much of this work possible. In particular, Columbus O'D. Iselin, director of the Institution, Adrian K. Lane, master of the Atlantis, and Dr. J. B. Hersey deserve special mention because of their unstinting co-operation. The National Geographic Society and the Bureau of Ships, Navy Department, as well as Columbia University, financed the work. The U. S. Navy Hydrographic Office contributed a portion of their data, thereby helping greatly in the progress of this study.

The Continental Slopes

In many areas the continental slopes exhibit appreciable changes in gradient. There is, for example, some evidence of flattening below 1600 fathoms off the coast of Africa. Off the eastern coast of the United States the detailed surveys of the U. S. Coast and Geodetic Survey show broad gently sloping shelf-like structures between 1300 and 1700 fathoms and between 2100 and 2300 fathoms in the broad embayment whose head is occupied by the Hudson Canyon (Tolstoy and Ewing, 1949, Pl. 3, a). This continental slope is dissected by a system of submarine canyons, the pattern of which suggests to some a subaerial origin. This pattern was discovered by Veatch and Smith (1939). Their investigations were limited to water shallower than 1600 fathoms south of the Delaware and to water shallower than 1200 fathoms in the area of and east of the Hudson Canyon. Two cruises by the Atlantis (spring and summer 1949) have shown that the Hudson submarine valley extends seaward at least 200 miles beyond the limits of the Coast and Geodetic Survey's work and is identifiable as far as the edge of the flat basin floor at 2600 fathoms. It was followed down the continental slope by means of a close series of zigzag sounding lines to a depth of 2540 fathoms, at which point its relief became too slight to identify it positively. Several tribu-



DISTRIBUTION OF THE MAJOR SUBMARINE PLAINS IN THE SAME AREA, AND SOUNDING LINES



Thick lines and letters refer to profiles in Plates 3-6. Thin continuous lines represent lines with continuous fathograms or with very closely spaced discrete soundings (spacing less than 3 miles). Dotted lines represent discrete sounding lines. Note that in the drafting of the contours of Plate 1 the isolated soundings given on the H. O. 0955 and 0956 series were used.

Vertical rules, 2850-2950 fathom plains; diagonal rules, 2700-2750 fathom plains; horizontal rules 2600-2650 fathom plains.

RINE PLAINS IN THE SAME AREA, AND SOUNDING LINES USED IN THE DRAFTING OF PLATE 1

taries enter it at grade between 100 and 2350 fathoms. Its relief is greatest where it cuts into the steeper slopes such as the upper 1200 fathoms of the continental slope and the slope between 1700 and 2100 fathoms. In this latter slope it cuts a gorge 1800 feet deep and roughly 3 miles wide in the ocean floor. In the almost flat areas between 2250 and 2300 fathoms it exhibits evidence of meandering. However the author does not believe that these features are sufficient in themselves to prove that the canyon had been cut out under subaerial conditions. A detailed presentation of the data on the Hudson submarine valley by Heezen, Ewing, and the author will be given elsewhere. On the bathymetric sketch of Plate 1 the small indentation in the 2500-fathom contour west of Caryn Peak represents the crossing of this contour by the Hudson Submarine Valley.

There are indications of other submarine valleys—such as the Hydrographer, Oceanographer, and Lydonia canyons—at depths greater than 2000 fathoms that prolong some of the larger submarine valleys of the American Continental slope. There are also several broad valleylike depressions in depths of 2500 fathoms in this area. The excellent control available here (Pl. 2) leaves little doubt as to the reality of these features. The continental slope northeast of the limits of the work of Veatch and Smith exhibits the same high degree of dissection; canyons occur down to depths of more than 2000 fathoms. South of Lat. 29°N., the continental slope becomes extremely steep (over 30° in some places) and suggests a fault scarp.

The Basins

Plate 1 illustrates the distribution of basins on both sides of the Mid-Atlantic Ridge. The nomenclature adopted is from Maurer and Stocks (1933). The floors of the North American, Northern Canary, and probably the Newfoundland basins are occupied mostly by vast smooth plains covering a total area of well over 200,000 square miles.

Thus the floor of the northwestern part of the North American Basin is characterized by a plain between 2600 and 2700 fathoms. In the area of the Caryn Peak its depth is 2600–2620 fathoms. It slopes very gently southeastward and southwestward; the average slopes are about 2 feet per mile. The northern boundary of this plain is clearly defined as the foot of the continental rise at 2600 to 2650 fathoms. The southern boundary is much less clearly defined but probably runs east-west, close to Lat. 35°N. Eastward it appears to merge into a somewhat narrower 2700–2750 level. The approximate location of these and other levels is shown in Plate 2. In some places the transition between the 2600–2650 and 2700–2750 levels seems to be steplike (Tolstoy and Ewing, 1949, Pl. 4, a). However, it is impossible in many cases to identify the boundary between these levels, since changes in depth of the order of 50 fathoms are very difficult to correlate from soundings by different ships or even different cruises of the same ship. The combined effects of small frequency changes in the instruments, of differences of sound velocity in the sea water, and of calibration may be of the order of 50 fathoms.

This 2700–2750 level merges eastward into the 2850–2950 plain occupying most of the floor of the northern section of the North American Basin. Both the 2600–2650 and the 2700–2750 levels appear to form narrow fringes along the foot of the continental rise west and south of the Grand Banks, between the rise and the 2800-fathom contour. The 2600–2650 plain may also occupy the southern end of the Newfoundland Basin.

The area covered by the 2850–2950 plain is shown in Plate 2. It slopes gently southward. Thus at Lat. 40°N. it lies at 2850 fathoms, at Lat. 37°N. it lies at 2890 fathoms, and at Lat. 31°N. its depth is 2950 fathoms. It has therefore a gradient of the order of 1 to 1½ feet per mile. Its southern end, near Lat. 30°N., is broken into a series of tonguelike flat-bottomed valleys trending northeast-southwest. These tongues deepen irregularly toward the southwest, sometimes reaching depths as great as 3300 fathoms. Their gradients are irregular and are frequently more than 10 feet per mile. To the west the 2850–2950 plain ends abruptly against a series of mountain ranges which mark the beginning of the Bermuda Rise. To the east, where it meets the foothills of the Mid-Atlantic Ridge, its boundaries are also well defined. The general character of this plain and of its boundaries close to Lat. 31°N. is illustrated by the profiles of Plates 4 and 5.

A similar plain at 2900 fathoms occupies a smaller area in the Northern Canary Basin. It

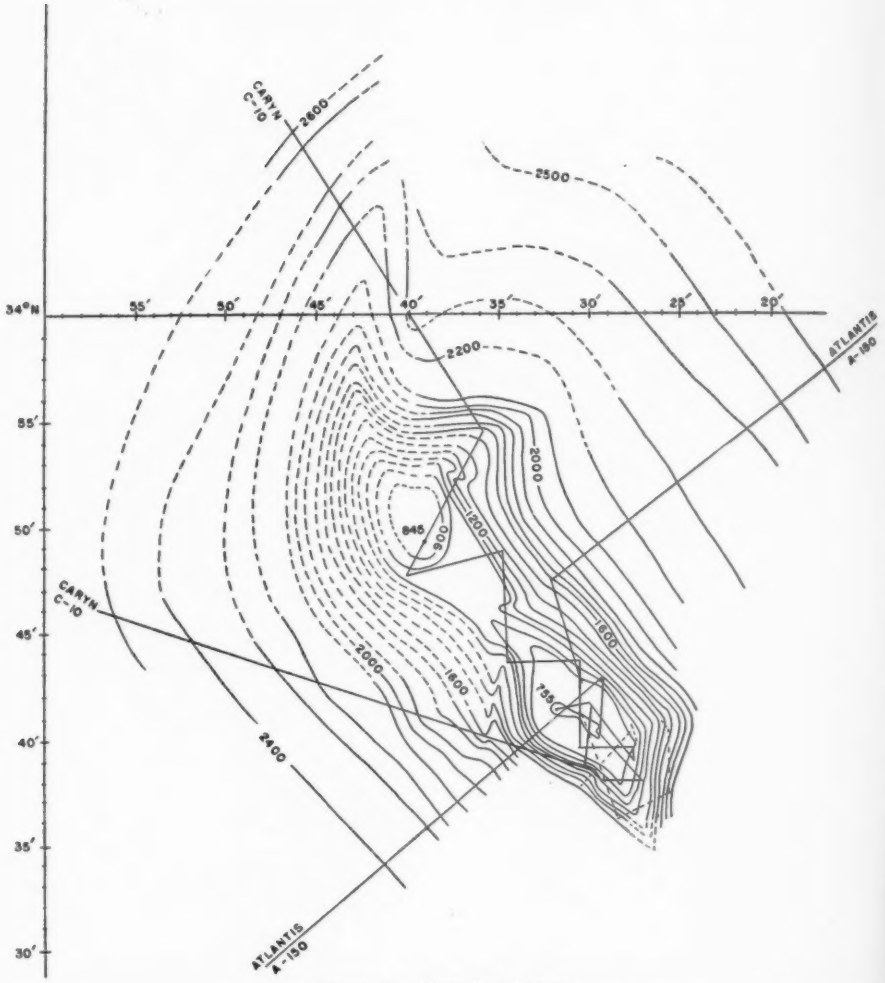
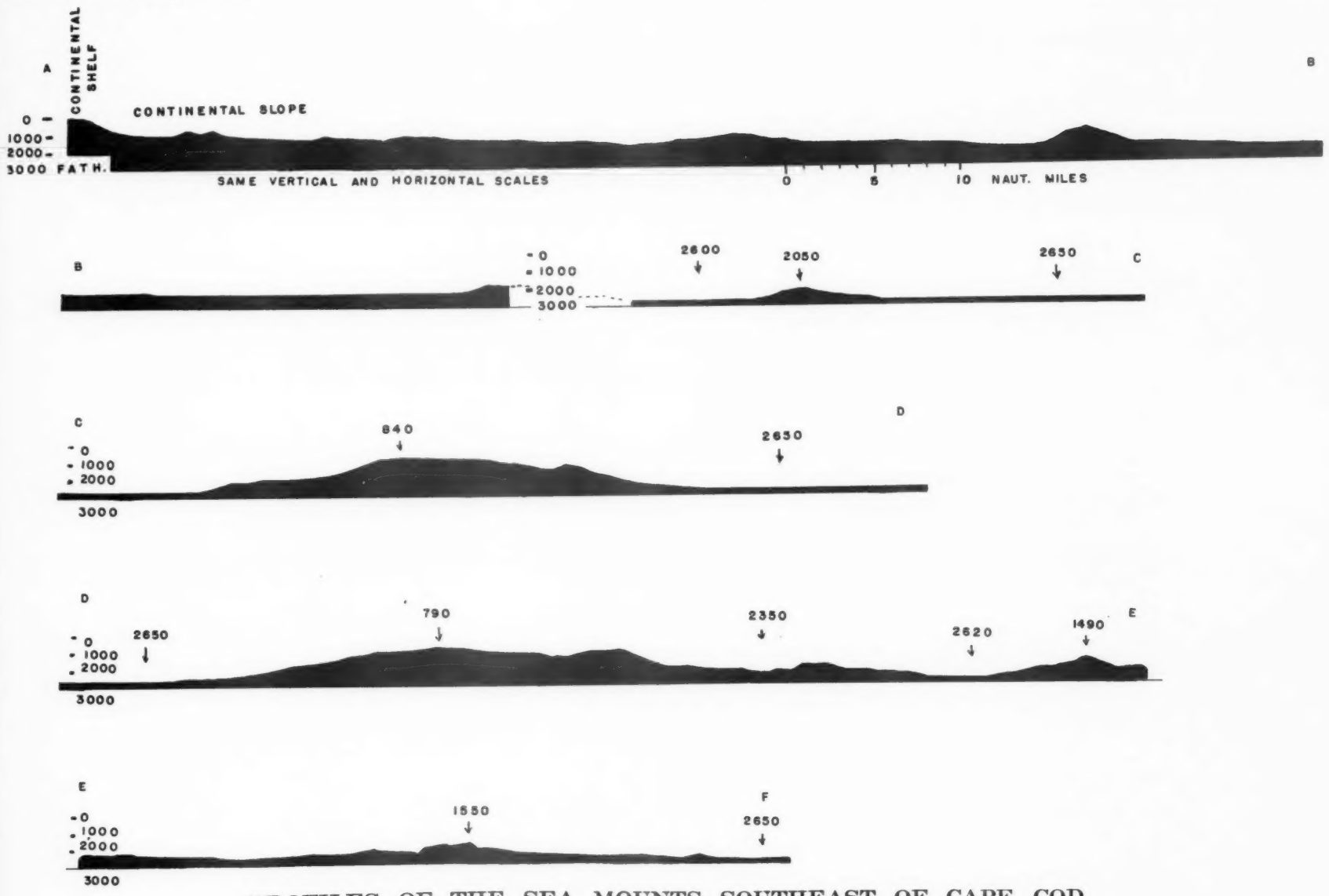


FIGURE 1.—MUIR SEA MOUNT

This is a new version of the sea mount shown by Tolstoy and Ewing (1949, Fig. 2), corrected by means of additional data obtained by the R. V. CARYN of the Woods Hole Oceanographic Institution during the summer of 1949. Contour interval 100 fathoms.

appears to be similarly broken up at its southern extremity into a series of tonguelike valleys deepening quickly but erratically to more than 3000 fathoms. A large gently undulating plain between 2800 and 2880 fathoms occupies the floor of the North American Basin southwest of Bermuda. It slopes gently southeastward.

Rising abruptly from these great gently sloping plains are numerous sea mounts of irregular sizes and shapes. Only a few have ever been investigated in detail. They vary in height from just a few fathoms to well over 9000 feet above the surrounding ocean floor. A particularly conspicuous group of flat-topped sea mounts rises from the 2600–2650 fathom plain, southeast of Cape Cod, to approximately 11,000 feet above the surrounding ocean floor (Pl. 1). They stand out conspicuously in Plate 1. A profile across



PROFILES OF THE SEA MOUNTS SOUTHEAST OF CAPE COD

Letters correspond to those in Plate 2. No. vertical exaggeration.

them is shown in Plate 3. Note that in this profile the vertical and horizontal scales are the same. There is a suggestion of terracing on their flanks, at a depth of approximately 2000 fathoms.

The Caryn Peak, at the mouth of the Hudson Submarine Valley, was investigated in some detail during the summer of 1949 by Ewing, Heezen, Northrop, and the author. Though the results of this investigation have not as yet been worked out in detail, Caryn Peak is a roughly conical peak, 1 mile high and 5 miles wide across its base; it rises abruptly out of the surrounding plain at 2600–2620 fathoms.

The sea mount, roughly 130 miles northeast of Bermuda (Tolstoy and Ewing, 1949, Fig. 2), was further investigated by the R/V CARYN in July 1949. It was named the Muir Sea Mount, in honor of the U.S.S. MUIR, which discovered it in 1945. A slightly revised and fuller map resulted from this investigation (Fig. 1).

THE MID-ATLANTIC RIDGE

The Mid-Atlantic Ridge occupies the central part of the North Atlantic Ocean, from Iceland to a point southwest of the Cape of Good Hope. The highest points of the Ridge rise to form islands such as St. Paul's Rocks, the Azores, and Iceland.

It may be subdivided into:

(1) A high central zone or the Main Range, which is shown on Plate 1 as a series of parallel ridges following the general trend of the Mid-Atlantic Ridge. Many of these ridges rise to less than 800 fathoms below the surface of the ocean. In several locations there are enough sounding lines to establish the pattern and to some extent the characteristics of these ridges, particularly between Lat. 30°N. and 34°N. and Long. 40°W. and 43°W. (Tolstoy and Ewing, 1949, Pl. 7). Plate 6 shows a profile across the Main Range in this area; the vertical and horizontal scales are in the ratio of 3.3/1. There appears to be a marked asymmetry of slope for many of these ridges; the eastward slope is usually the steeper.

(2) An intermediate zone flanking the Main Range to the east and to the west between 1600 and 2500 fathoms south of the Azores is characterized by a succession of levels which for lack of any better term have been called terraces. This zone has been named the Terraced Zone (Pls. 5, 6). There are two types of transition between successive terraces. Some are separated by a step, with no intervening peak, whereas in others there is a peak or hump on the outer edge of the higher terrace. Reflection-shooting studies by Ewing, Press, and others have shown that, while the Main Range itself is almost devoid of sediments, the terraces are characterized by important thicknesses of sediment, of the order of one or several thousand feet. Certainly therefore, these terraces are or have been areas of greater deposition.

(3) An outer mountainous zone, between the 2900-fathom plains of the Northern Canary and North American Basins and the Terraced zone, has peaks that seem to form roughly north-northeast—south-southwest ridges. This Zone resembles neither the Main Range nor the Terraced Zone types of topography and stands out as a distinct physiographic province. Some of its peaks are more than 3000 feet high. For this area the name of Foothills of the Mid-Atlantic Ridge seems appropriate.

Plate 5, a typical profile across the western Foothills, shows a high peak (1700 fathoms) at the boundary between the Foothills and the Terraced Zone. It is flanked by deeps. The western, or outer, deep with respect to the Mid-Atlantic Ridge may be as deep as 3050 fathoms and has a flat bottom. The eastern, or inner, deep is 2550 fathoms deep, has a flat bottom, and marks the beginning of the Terraced Zone. Since we are ignorant of the processes that formed the Terraces and this peculiar type of feature, we cannot say whether it would be consistent to identify this flat stretch at 2550 fathoms as the first terrace.

The whole length of the boundary between the Foothills and the Terraced Zone is characterized by an essentially continuous ridge shoaling many places to appreciably less than 2000 fathoms and flanked by troughs (Pl. 1). The western trough is the deeper; it reaches depths greater than 3000 fathoms in several places, whereas the eastern one is commonly 2500–2600 fathoms deep and in some places less. The continuity of these troughs would be demonstrated best by tracing the 2800- and the 2400-fathom contours.

Outstanding Physiographic Features

The most striking feature is the Mid-Atlantic Ridge itself. In this area its general trend is northeast-southwest south of the Azores and northwest-southeast north of the Azores. It therefore roughly parallels the North American and African coasts. This fact is true of most of the Ridge and has long been pointed out in the literature. Thus Prinz (1891) considered it as evidence in favor of his hypothesis of the differential torsion of the hemispheres, and Wegener (1924) considered that it favored his theory of continental drift.

Close examination of the Ridge indicates that upon this trend are superposed transverse ones. Thus we find the deep trough surveyed by the Atlantis on its first cruise across the Mid-Atlantic Ridge in 1947 (Tolstoy and Ewing, 1949, Pls. 7, 8), which trends east-west along Lat. 30°N. and which has some characteristics of a graben.

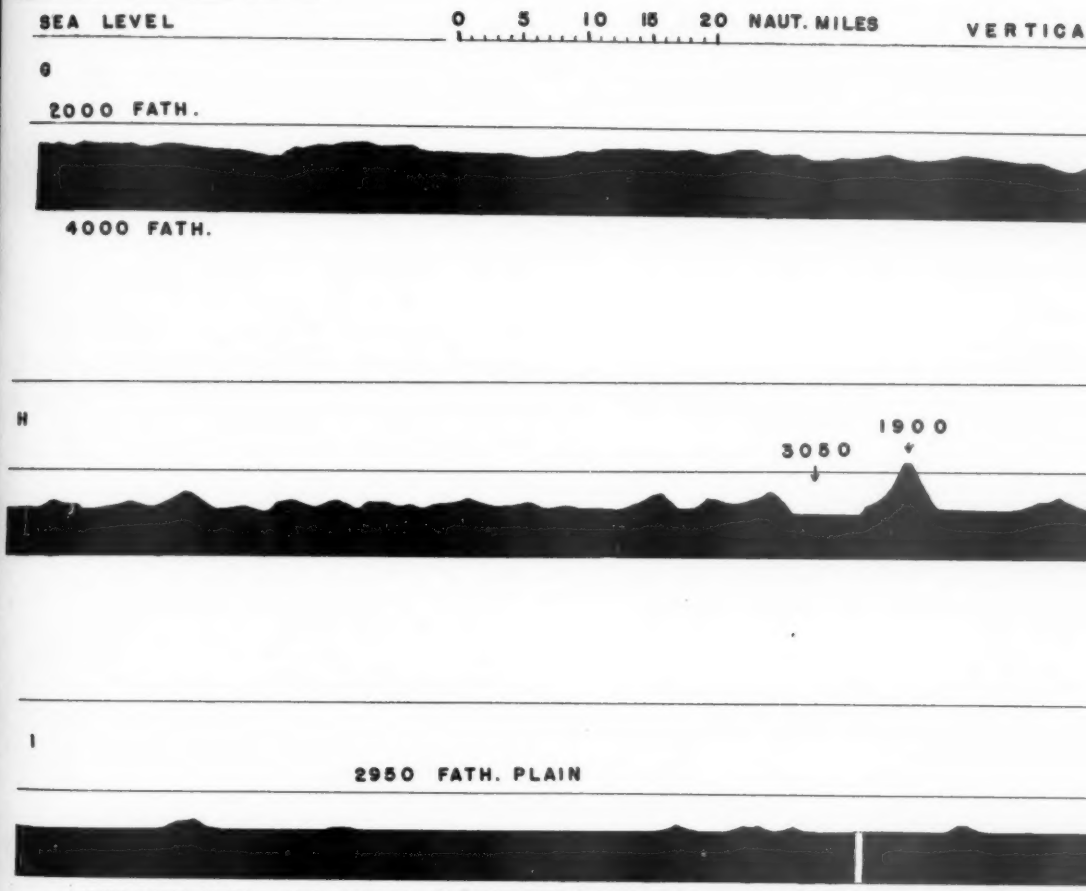
The Azores Bulge and its ridges are characterized by predominantly northwest by west—southeast by east trends. Later data obtained by the Atlantis and other vessels have resulted in a corrected version of Wüst's map (Fig. 2). The main control lines are shown in Figure 3. For the isolated soundings utilized, see Wüst's original map (Wüst, 1939). This more up-to-date interpretation shows indications of a north-south trend, practically at right angles to the preceding. This north-south trend would represent the true Mid-Atlantic Ridge trend in this area if we imagined the Azores to be absent and the Mid-Atlantic Ridge as bending around, changing its trend from northeast-southwest south of the Azores to northwest-southeast north of the Azores. The transverse trend would either have been superimposed upon or superimposed by the Mid-Atlantic Ridge trends. This idea is further enhanced by the fact that the transverse trend of the Azores appears to be a local manifestation of a major transatlantic one, beginning at the Grand Banks, crossing the Ridge at the Azores, and going through the Gettysburg and Josephine banks and merging with either the Variscian mountains of western Spain, Portugal, and North West Africa, or the Cenozoic ranges of southern Spain and North West Africa. Since the structure in southern Newfoundland and in the Prince Edward and Cape Breton islands is mostly Acadian and Appalachian, the Grand Banks and indeed this whole transatlantic trend may be Variscian.

The chain of sea mounts extending southeastward from Cape Cod form another transverse trend which can be followed up to the ridge marking the boundary between the Foothills and the Terraced Zone of the Mid-Atlantic Ridge. At Lat. 36°40′ N., Long. 55°W. a sea mount rises more than 200 fathoms above the 2900-fathom plain (Pl. 2). This trend is more continuous therefore than one would infer from a glance at Plate 1, the contour interval of which is such as not to show this sea mount. The highest point yet found on the ridge marking the boundary between the Foothills and the Terraced Zone is in the general area of its intersection with this trend.

The northeast-southwest trend of the ridges and valleys at the southern border of the great 2850–2950-fathom plains in both the North American and Northern Canary basins is interesting and is strikingly illustrated by the 3000-fathom contour (Pl. 1). In the Northern Canary Basin, control is poor, but the picture is believed to be at least qualitatively correct.

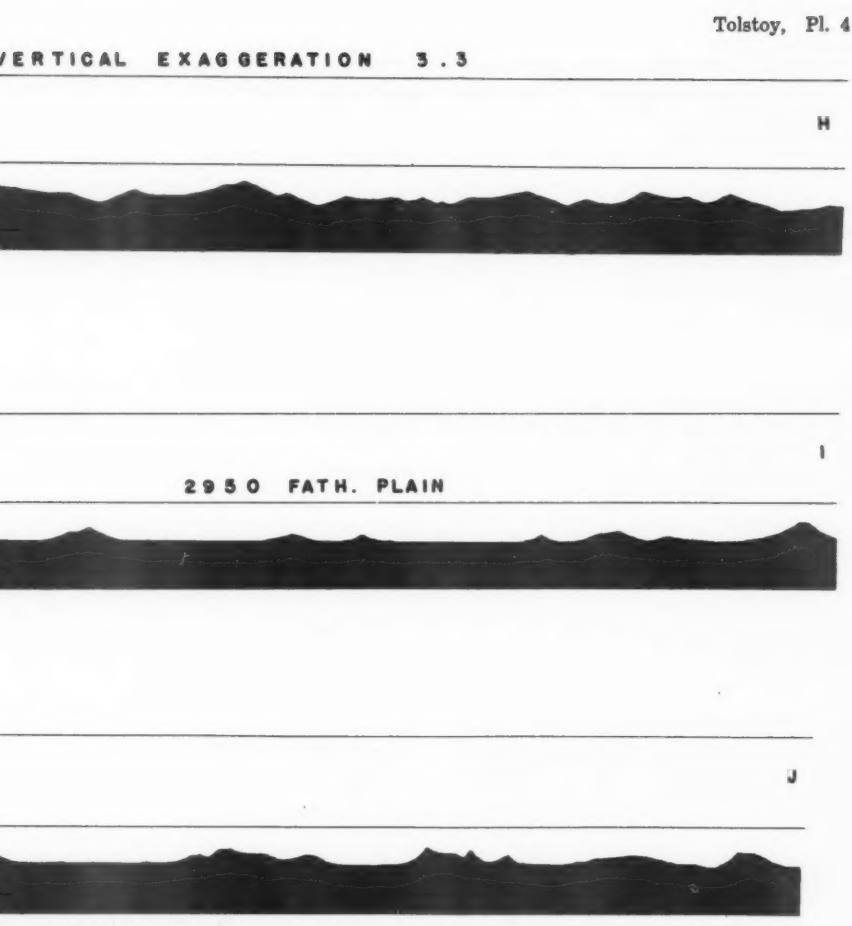
Along the southern rim of the 2850–2950-fathom plain in the North American Basin, the control is much better (Pl. 2), particularly in the section immediately adjacent to the Bermuda rise. The first indications of this type of feature may be noted northwest of Bermuda. It becomes much more pronounced southeast of Bermuda. At least as far as Long. 56°W. there is enough control to insure that broad features shown are correct, though of course the perfect continuity of these ridges and valleys may be open to doubt.

The parallel, closely spaced ridges so typical of the Main Range southwest of the Azores are not so pronounced north of these islands. There appears to be in that area more isolated peaks, which could well be volcanic in their origin; the Terraced Zone is no longer as pronounced, and the distinction between the three types of topography not so obvious. However, the number of sounding lines with continuous graphical records crossing this section of the Ridge is small, and there is no means of telling how general this change in character might be.



PROFILES OF THE SOUTHEASTERN EDGE OF THE BERM
FATHOM PLAIN, CLOSE TO THE 31ST

Letters correspond to those in Plate 2.



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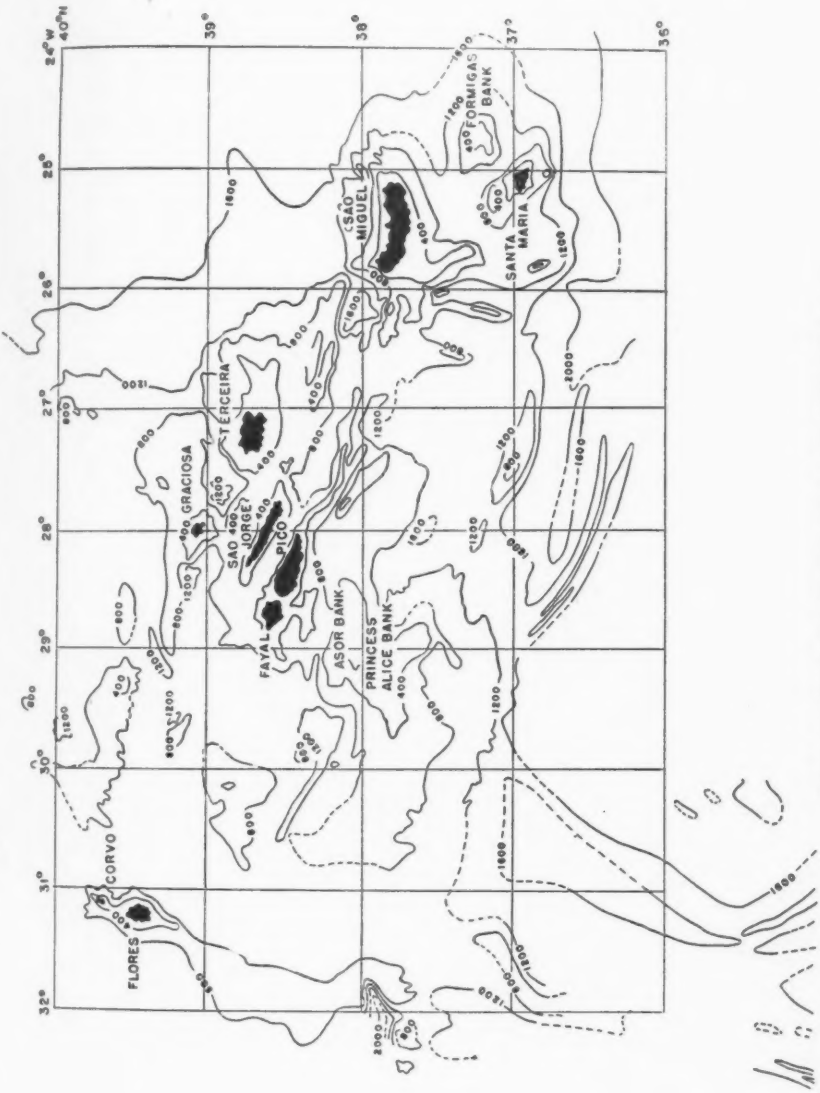

FIGURE 2.—SUBMARINE TOPOGRAPHY OF THE AZORES BULGE

This is a corrected version of Wüst's map, revised in the light of later information. The contour interval is 400 fathoms.

Outstanding Physiographic Features

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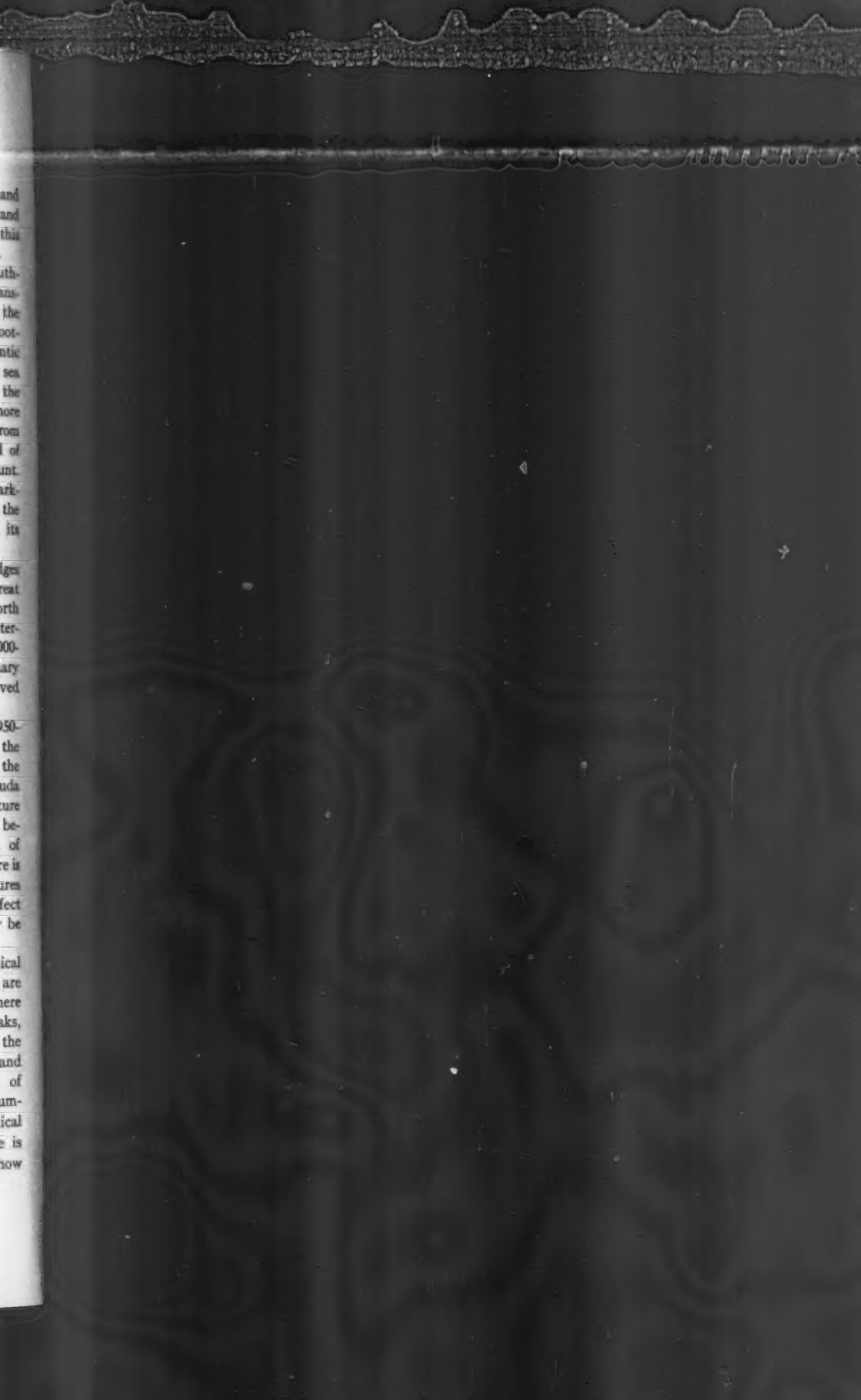
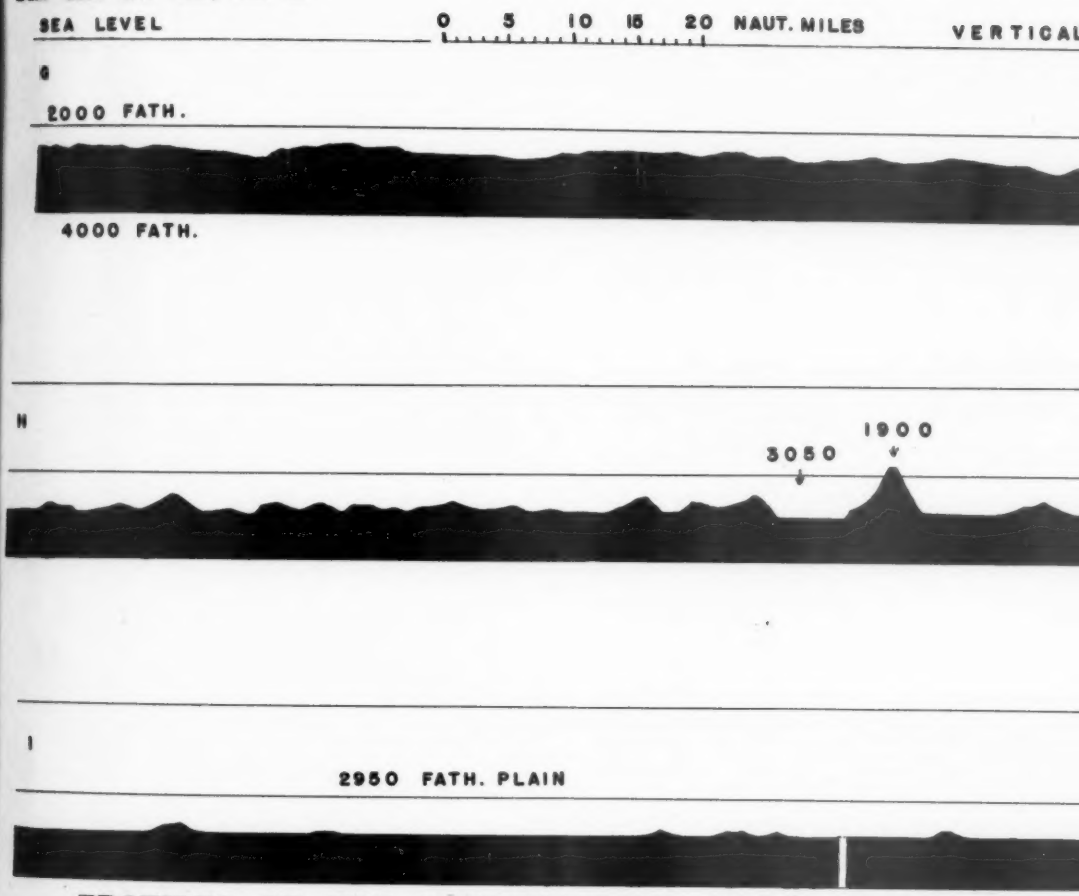
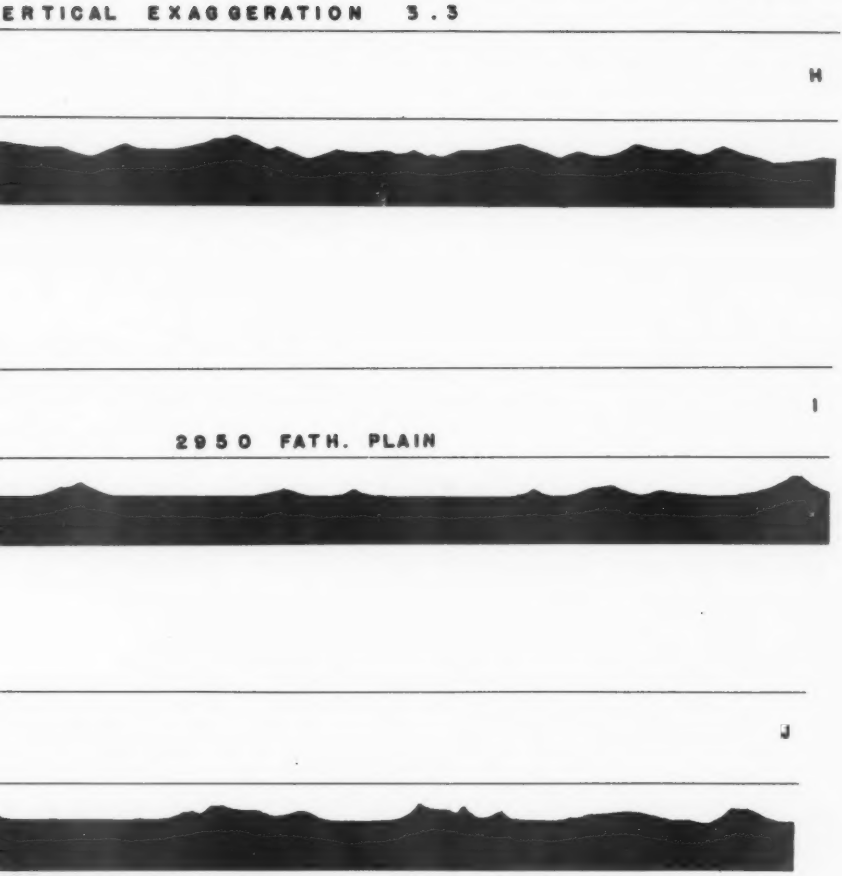




PROFILES OF THE SOUTHEASTERN EDGE OF THE BERM
FATHOM PLAIN, CLOSE TO THE 31ST

Letters correspond to those in Plate 2.



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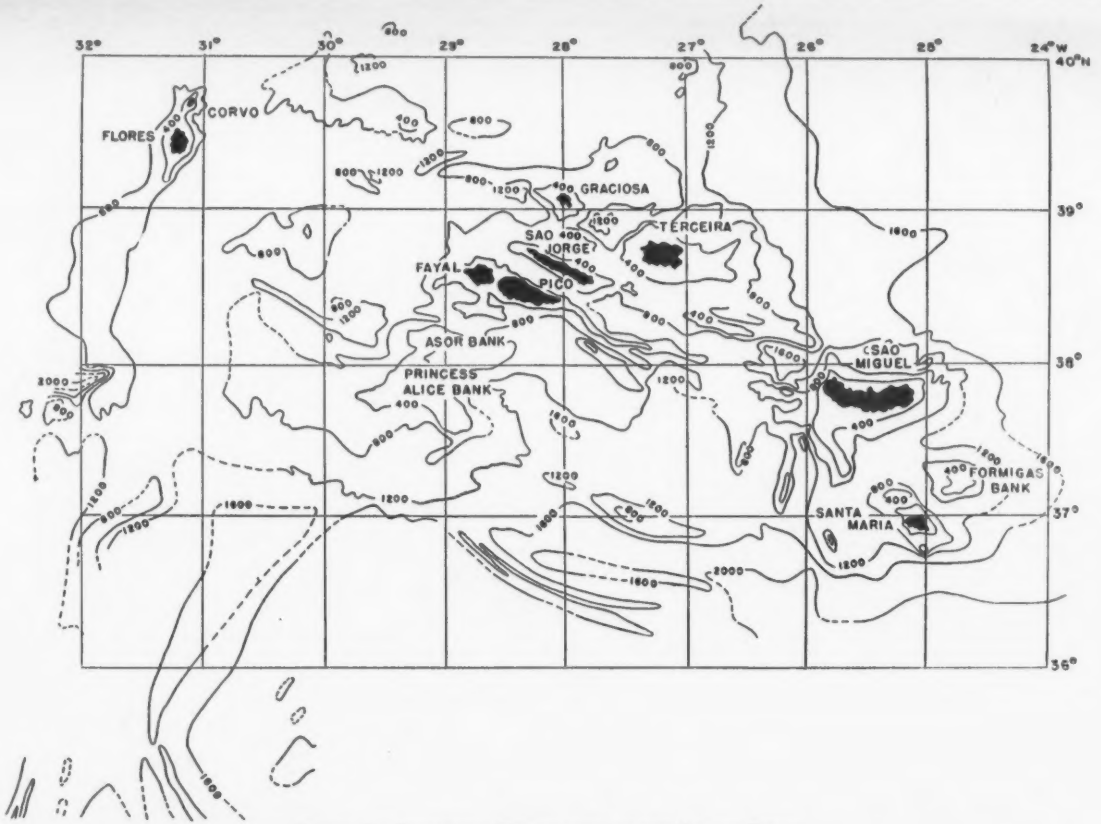

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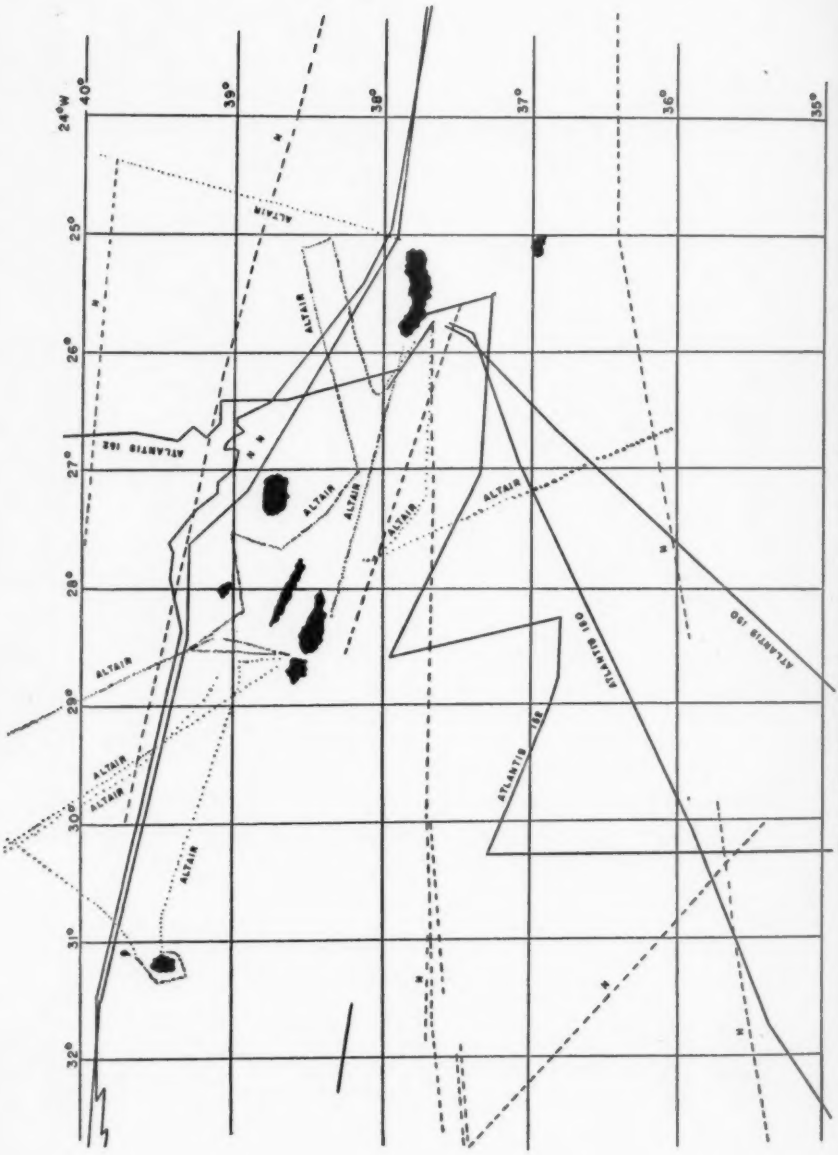

FIGURE 3.—SOUNDING LINES UTILIZED IN THE MAP OF FIGURE 2

Solid black lines indicate tracks with continuous fathograms, four of which were obtained by the ATLANTIS. The principal ALTAIR lines are reproduced. For the isolated soundings utilized in this interpretation, *see* Wüst's original map and local Portugese charts. Tracks marked N are U. S. Navy Hydrographic Office sounding lines belonging to the series already utilized in a previous paper. Solid black areas are islands (*See* Figure 2.)

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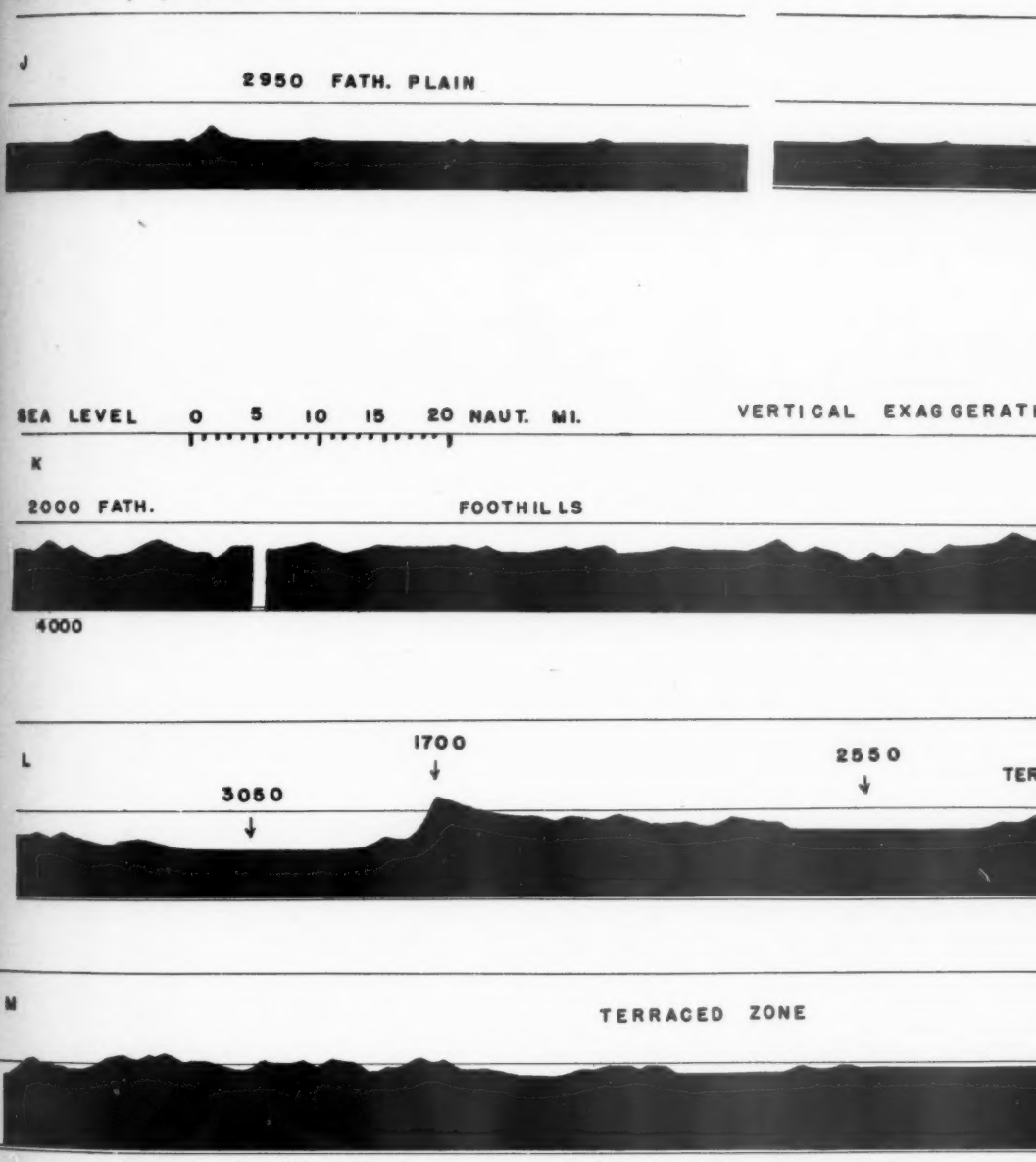
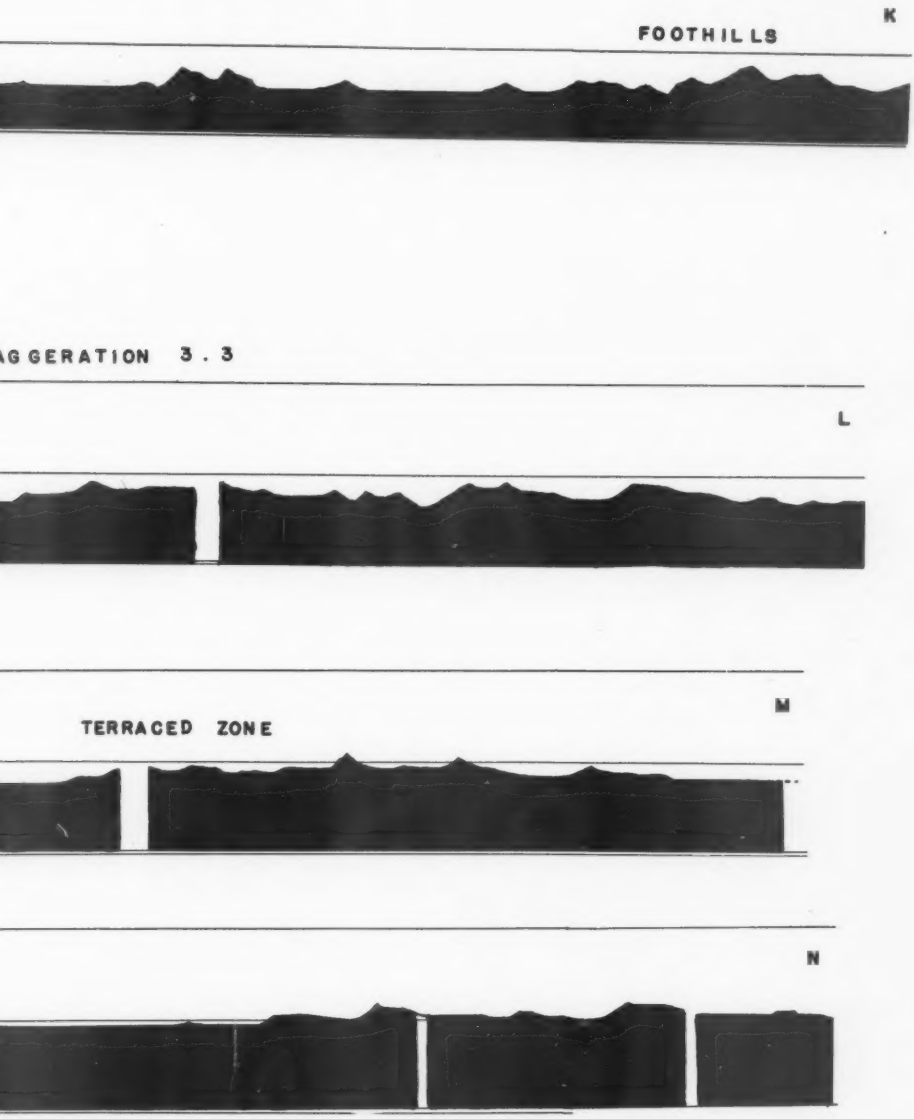


PROFILES ACROSS THE EASTERN EDGE OF THE 285
PARALLEL, THE WESTERN FOOTHILLS, AND THE TERRAC

Letters correspond to those in Plate 2.



HE 2850-2950 PLAIN CLOSE TO THE 31ST
ERRACED ZONE OF THE MID-ATLANTIC RIDGE
e in Plate 2.

Finally, there is a striking arcuate submarine mountain range branching out from the Mid-Atlantic Ridge southwest of the Azores which extends through Cruiser Bank. The sea mount described by Tolstoy and Ewing (1949, Fig. 4) which shoals to 180 fathoms is part of this mountain range. One cannot say whether the Great Meteor Bank, 130 miles south-southeast of Cruiser Bank, is part of this range.

Comments on the History and Structure of the Mid-Atlantic Ridge

The question as to whether the Mid-Atlantic Ridge owes its present character to folding or faulting, or to a combination of both, cannot be answered. That there is considerable evidence of fracturing on the Ridge is indisputable. Thus the east-west trough cutting through the Main Range at Lat. 30°N. is almost certainly a fault, probably a graben. That the northwest by west—southeast by east ridges on the Azores Bulge may be due to the outpouring of magma through fissures opened during the doming of a section of the earth's crust, as proposed by Hans Cloos (1939), is quite possible. The central part of Iceland is probably a graben (Hawkes, 1938; 1941).

The structure characterizing the Main Range could have been produced by block faulting. W. H. Bucher pointed out to the author that this type of topography is not essentially different from many fault-controlled mountain ranges on the surface of the earth, such as those of Southern Oregon and of East Africa. However, this type of topography could have been equally well produced by folding. This idea is not in contradiction with any known facts concerning this section of the Mid-Atlantic Ridge, particularly since samples of consolidated sedimentary rock have been brought to the surface by dredging. Of course, it remains to be seen how much of the Mid-Atlantic Ridge is actually composed of such rocks. Furthermore, if the structure of the Azores Bulge owes its present character to the superposition of two separate structural processes, the mechanism of the formation of northwest by west—southeast by east ridges as proposed by Hans Cloos might have to be revised.

At least this much may be said of the area of the Azores: it is not characterized by typical Mid-Atlantic Ridge type of structure, but probably owes its present character to the combined effect of at least two major structural events, and therefore does not constitute a typical "outcrop" of the Mid-Atlantic Ridge. Whatever may be said of the structural history of this area does not therefore necessarily apply to the rest of the Ridge. Unfortunately the same may be said to a certain extent of Iceland, where the structure is governed by block faulting accompanied by huge outpourings of basalt. It is also situated at the intersection of the Mid-Atlantic Ridge with a transverse trend, sometimes interpreted as Caledonian. The major faults bounding the central Icelandic graben strike roughly northeast-southwest— *i.e.*, parallel to the Mid-Atlantic Ridge southwest of Iceland.

There is a rough and admittedly unsatisfactorily established synchronization between the periods of major volcanic activity in Iceland and in the Azores which can perhaps help us estimate the times at which major structural upheavals may have occurred in the history of the Mid-Atlantic Ridge. Hawkes has shown that the major outpourings of basalt in Iceland (the so-called Thulean basalts) must have occurred in pre-Miocene time. Agostinho (1936) on the other hand believes there is evidence of a pre-Miocene period of volcanic activity in the Azores. Also in both Iceland and the Azores there has been considerable post-Miocene and recent volcanic activity.

There are thus two major periods of activity evident in the history of the Azores and Iceland. One of these would be late Tertiary to recent, the other middle or early Tertiary. This suggests that the origin of the Ridge might be associated with the Alpine orogenies. This idea is supported also by the relatively high degree of seismicity along the Ridge, which is comparable to that of many sections of Gutenberg's Alpide Zone. Gutenberg and Richter (1949) have even tentatively suggested that the Mid-Atlantic Ridge may be a belt of Tertiary folds, the high degree of seismicity of which would be due to block faulting following a redistribution of tectonic forces.

Furthermore, the processes that produced the Mid-Atlantic Ridge may have been superim-

posed upon older transverse trends, possibly Paleozoic, resulting in the opening up of fractures through which huge masses of lava were extruded. At these points one notes a higher degree of seismicity.

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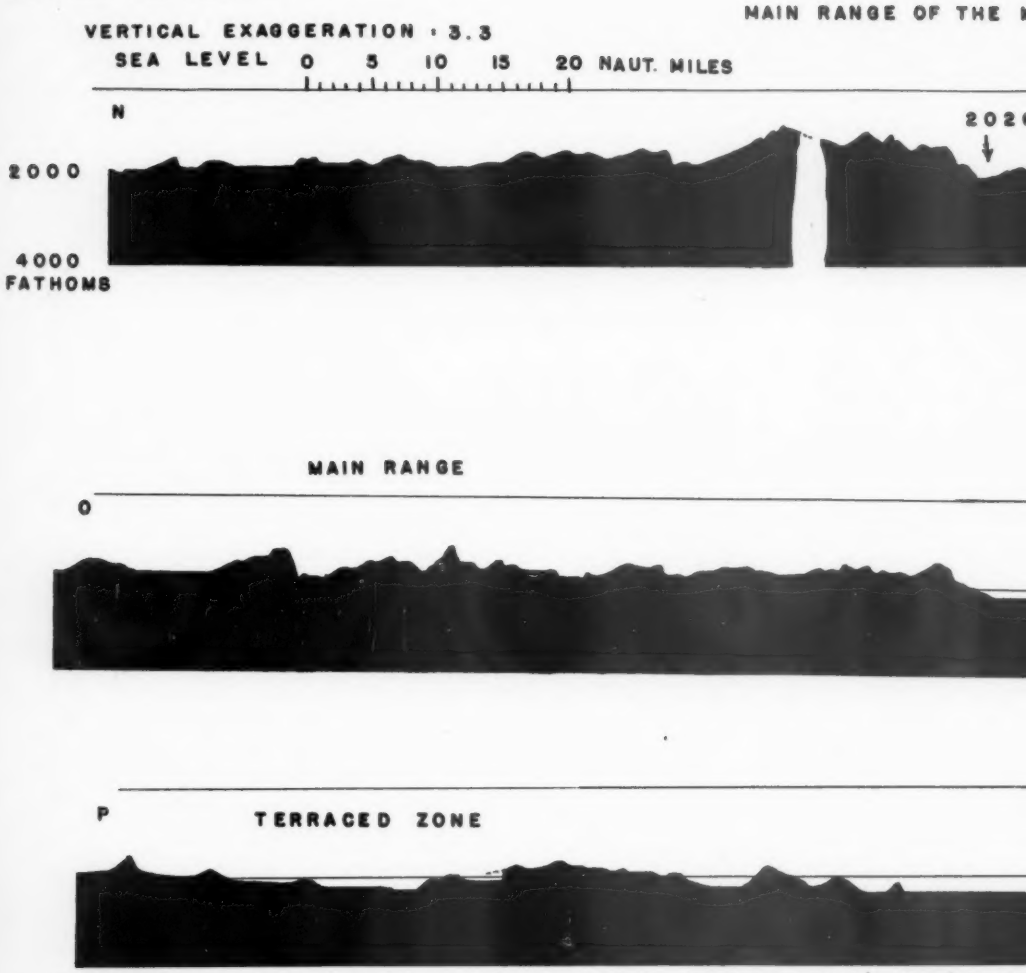
Contribution No. 559, the Woods Hole Oceanographic Institution.



PROFILES ACROSS THE TERRACED ZONE AND THE MA

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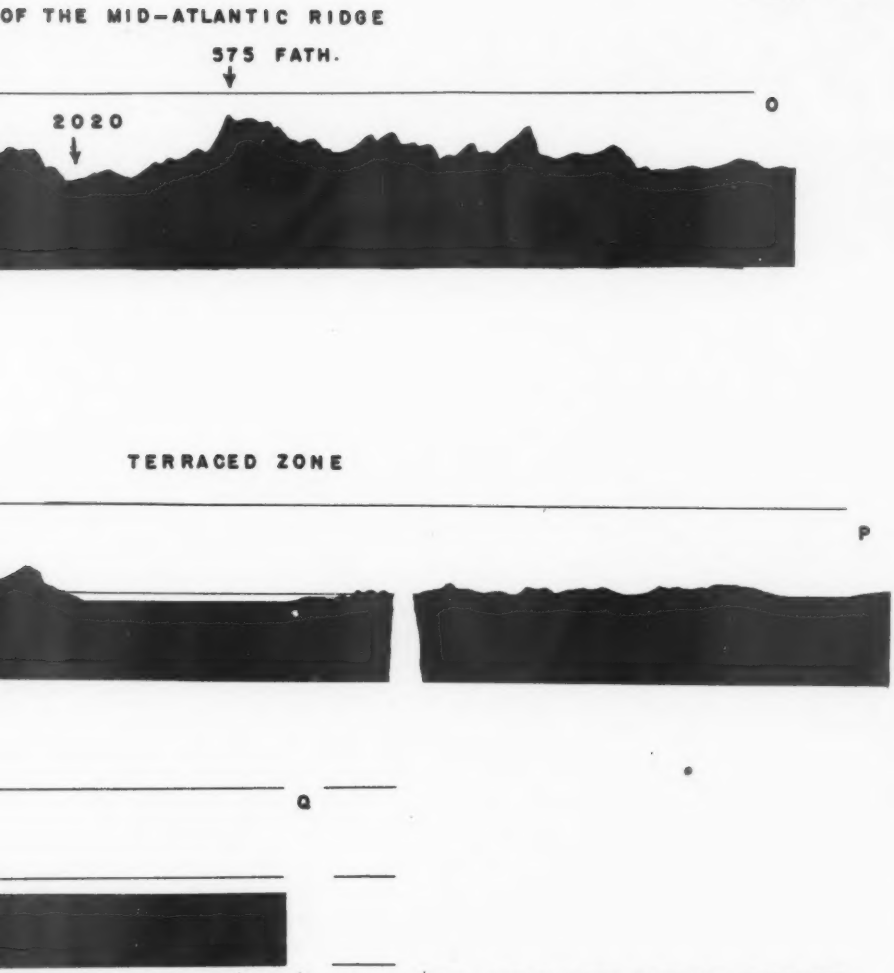

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ALTERATION FEATURES AT SILVER BELL, ARIZONA

By PAUL F. KERR

ABSTRACT

The Silver Bell Range, some 50 miles west of Tucson, Arizona, provides a varying array of igneous and sedimentary rocks in an area of widespread copper mineralization. Two deposits have been explored, Oxide on the south end of the Range and El Tiro 3 miles northwest. Both are essentially disseminated deposits and exhibit a series of hydrothermal metamorphic stages superimposed upon a variety of geologic conditions.

Early mined copper ores were largely formed by contact-metamorphic action along the boundaries between dacite porphyry or quartz monzonite and Paleozoic limestones. The contact deposits in the Mammoth-Union area at Silver Bell and pockets at El Tiro reportedly produced 100,000,000 pounds of copper. Disseminated ores explored in this investigation differ from the contact ores in being largely associated with quartz monzonite and to a minor extent dacite porphyry or dacite.

Major deformation has resulted in faulting of Precambrian alaskite on the southwest against Paleozoic sediments on the northeast. Major post-Paleozoic igneous activity has involved a dacite flow, a dacite porphyry intrusive, and quartz monzonite in sequence. Minor intrusive activity is represented by a series of dikes invading previously formed rocks and frequently following fault lines.

Hydrothermal copper-depositing solutions which followed the quartz monzonite have produced a halo of alteration minerals in alaskite, quartz monzonite, and the two dacites. The effects are observable chiefly in the quartz monzonite, but with equal intensity in limited areas of the other igneous rocks. Convergence in metamorphism with ultimate quartz-sericite mass replacement may be observed in all four rocks.

The alteration minerals are kaolinite, halloysite, montmorillonite, illite (or hydromica), sericite, chlorite, alunite, jarosite, barite, opal, secondary quartz, and adularia. Field criteria confirmed by laboratory study have been used to divide the four igneous rock types into alteration stages. An alteration map of the entire mineralized area shows these stages, and significant portions of the map are reproduced in the paper.

Certain aspects of the alteration at Silver Bell, Arizona, agree with stages described by Sales and Meyer (1948) at Butte, Montana; Lovering (1949a) in the East Tintic area, Utah; Peterson *et al.* (1946) at Castle Dome, Arizona; and Kerr *et al.* (1950) at Santa Rita, New Mexico. Where the early rock contains ferromagnesian minerals a chloritic stage is prevalent. Plagioclase is more susceptible to argillic alteration than orthoclase although both succomb. Kaolinite and illite (or hydromica) are intermediate. Sericite and quartz predominate toward the end of the hydrothermal stage at or near convergence. With the possible exception of East Tintic there also appears to be a striking parallel of the argillic stage with metallic mineralization. Silver Bell and Santa Rita agree closely in this respect.

Metallic mineralization appears to coincide with areas of advanced argillic alteration. This is shown most clearly in the quartz monzonite, the chief ore-bearing rock. Alteration zones followed on the surface aid in outlining areas for drilling and in excluding large unmineralized areas from consideration.

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INTRODUCTION

Objectives

In the western portion of the Silver Bell Range, Arizona (Fig. 1), the writer has studied the relationship between hydrothermal alteration and disseminated copper mineralization. Although the mineral deposits are not large in comparison with a number of copper-bearing porphyry areas of the west, several features are extraordinarily significant.

Of greatest interest is the Silver Bell alteration pattern. Four igneous rock types, each with an individual alteration sequence, culminate in metamorphic convergence. The progressive alteration from freshly crystallized rocks to an ultimate clay mass also provides a proving ground to co-ordinate field with laboratory study and to evaluate criteria.

The two disseminated copper deposits of Silver Bell, Oxide and El Tiro, provide examples of highly altered areas where ore bodies are known but are still unmined. Since portions of the area have been explored both underground and by drilling, wall-rock alteration can be correlated with sulphide mineralization where both surface indications and knowledge of the ores below are available.

The relationship between structure and alteration and the system of dikes following the belt of thrust faulting is worthy of attention but is beyond the scope of this study. Such geological problems as the history of the contact metamorphism and the stratigraphic relationships have been passed over briefly in order to concentrate on the alteration problem.

Recent studies of wall-rock alteration in the vicinity of ore deposits include those of Lovering and associates (1949a; 1949b) in the vicinity of Eureka, Utah; Sales and Meyer (1948) at Butte, Montana; Peterson *et al.* (1946) at Castle Dome, Arizona; Schwartz (1947) at San Manuel, Arizona, and elsewhere; and Kerr, Kulp, Patterson, and Wright (1950) at Santa Rita, New Mexico.

All these studies are concerned with the significance of clay and other finely crystalline material in association with ore formation. All apply particularly to the ore-forming processes occurring as an aftermath of igneous invasions

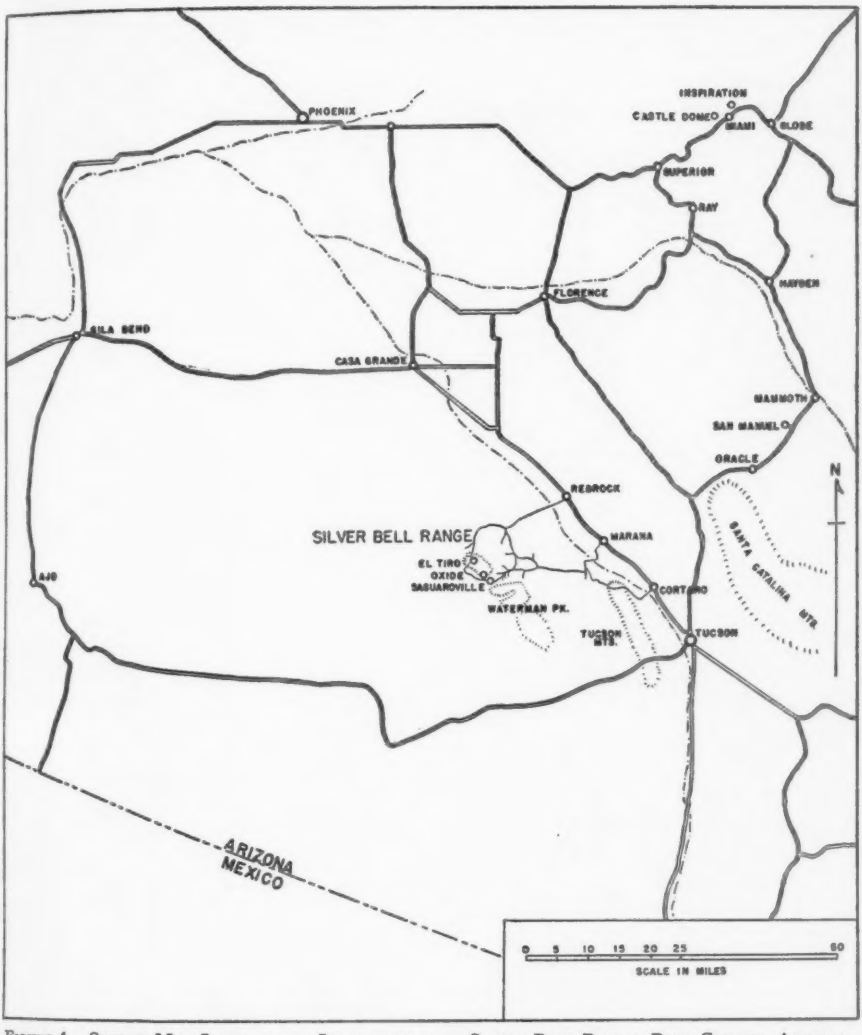

FIGURE 1.—SKETCH MAP SHOWING THE LOCATION OF THE SILVER BELL RANGE, PIMA COUNTY, ARIZONA

of the Cordillera of the western United States. All are concerned much more with the broad processes of hydrothermal alteration than with later supergene processes. All have found significant zoning emphasized particularly by distribution patterns of such minerals as kaolinite, sericite, hydromica, and montmorillonite. All tend to emphasize the role of clay minerals and other fine materials as indicators of the trends of sulphide deposition.

While such areas of general agreement point to substantial progress in the development of another method applicable to the search for the so-called hidden ore body, more experience will be required before these studies may be fully evaluated and generally applied as a routine technique. Too few data are available directly correlating known ore zones with carefully determined alteration zones, and the geochemical significance of the alteration minerals is too im-

perfectly understood. The old problem of the distinction between hydrothermal and supergene phases is also worthy of attention. Likewise too few data are available in correlating mineralization with alteration intensity to draw broad generalizations.

Location

The Silver Bell Range lies about 40 miles northwest of Tucson (Fig. 1) with the abandoned mining camp of Silver Bell at the north end. The group of hills, ridges, and peaks which constitutes the range is roughly 5 miles long and 3 miles wide, with one disseminated copper deposit, El Tiro, at the northwest end, and a second, Oxide, at the southern end (Pl. 1).

Early operations centered around the Mammoth and Union mines along the contact between quartz monzonite and Paleozoic limestone at the eastern and southern side of the Silver Bell camp and the Hilda, Lauzanne, Daisy, and Kurtz mines along a group of fissures to the west. Mining activity in the area dates back to 1863, and records indicate a total production of around 100,000,000 pounds of copper up to 1930, when mining was discontinued.

An area of disseminated copper deposition in the vicinity of the abandoned Hilda, Lausanne, Daisy, and Kurz mines is now generally referred to as the El Tiro deposit. A somewhat larger area at the opposite end of the range constitutes the Oxide deposit. Alteration studies have been chiefly concerned with these two bodies.

Silver Bell was once connected with the Southern Pacific Railroad at Red Rock by the Arizona Southern Railroad which is now merely an abandoned road bed. The copper ore was smelted at Pasco (Barney, 1904).

According to Shaw (1909) the principal mines worked in the early days at Silver Bell were the Mammoth, Union, and Billy. These removed copper of 3- to 4-per cent grade largely associated with garnet.

The town of Silver Bell is now largely stripped to a group of abandoned foundations, although, as pictured by Stewart (1912), it was a respectable mining community.

Procedure

The program of alteration study has consisted of a combined field and laboratory investigation. Emphasis has been placed upon the role of the clay minerals as indicators of the extent of hydrothermal alteration of the rock types associated with copper mineralization. For mapping, field criteria were first developed to indicate the progress of alteration. Study in the laboratory included the examination of thin sections, the optical examination of fragments, thermal analysis, and X-ray diffraction technique.

Participation in active field work began on February 9, 1948, and continued until June 3. Throughout the entire period Mr. J. C. Playter, resident geologist for the Trench mine, assisted in the field, and Miss P. K. Hamilton assisted with laboratory investigations.

Results of an extended geological survey of the district prepared by H. M. Kingsbury, I. P. Entwistle, and Harrison Schmitt (1941), and representing 9 months of careful geological work, have been available. A set of aerial photographs by Fairchild Aerial Surveys, Inc., and a complete set of topographic maps on scales of both 400 and 200 feet to the inch have also been available. Prior to this work a modern camp was established at the south end of the Silver Bell Range, and access roads were opened, making it possible to drive within a reasonable working distance of all parts of the area.

When the field work was well underway, active churn drilling was begun on the Oxide deposit, and a program of sampling was initiated under the direction of Mr. Harold Courtright and Mr. Richard Mieritz. Company studies undertaken in connection with this drilling program have been of considerable assistance.

Acknowledgments

The co-operation of the American Smelting and Refining Company in supporting the research project and in granting permission to publish the results is greatly appreciated.

The writer is also indebted to Mr. A. E. Ring, formerly manager of the Southwestern Division of the Western Mining Department of Asarco[1], for following the work with interest

[1] The American Smelting and Refining Company is frequently referred to as Asarco.

Atlas mine
JESUIT HILL
SILVERBELL CANYON
QUARTZITE HILL
Deep shaft
Wilson mine
Kurtz mine
Daisy mine
Union mine
HOPE CANYON
Williams mine
Lauzonne mine
Hilda mine
DACITE
TIGRO CANYON
Mammoth mine
MT. EXPECTATION
MT. FALTO
ALASKITE
MAMMOTH CANYON
FOURTH
MONZONITE
QUARTZ MONZ
LEGEND
LATE DIKES
QUARTZ
QUARTZ MONZONITE PORPHYRY
EARLY DIKES
DACITE PORPHYRY
DACITE
ALASKITE
PERMIAN? BROWN SANDSTONE
PERMIAN? RED BEDS
PALEOZOIC SEDIMENTS
0 1000 2000 3000 4000
SCALE IN FEET

OF THE SILVER BELL AREA, ARIZONA

and providing facilities both in the Tucson office and in the field, which made the work both effective and enjoyable. Dr. Walter Landwehr, chief geologist, devoted several days in the field to the discussion of the alteration problems and the interpretation of field criteria, paying particular attention to applications of the work. Mr. Kenneth Wilson, resident geologist, has assisted through the discussion of problems of geologic interpretation both in the field and office. Mr. F. M. Stephens of the Tucson office has aided in providing data from extensive files on Silver Bell.

Mr. Harold Courtright, geologist in charge of the Silver Bell drilling program, and Mr. Richard Mieritz, his engineering associate, have been particularly helpful throughout the field work. Their constant contact with the practical problems of exploration has helped greatly in focusing the attention of the alteration study on its ultimate application.

In particular, Mr. John C. Playter of Asarco and Miss P. K. Hamilton of Columbia University have contributed extensively to the investigation.

Geological Relationships

General Statement

Because of the study of the geology of the Silver Bell area by Kingsbury, Entwistle, and Schmitt (1941), discussion of the general geology is here limited to an outline of the geological features pertinent to the alteration study. The general geologic relationships are shown in Plate 1. Silver Bell furnishes a wide variety of geological problems beyond the range of this study involving in particular stratigraphic relationships, contact metamorphism, and structure. It has been impossible to give these problems more than passing notice.

The formations from Cambrian to Permian (Fig. 2) have been considered as a group without attempting separate mapping of the Bolsa, Abrigo, Martin, Escabrosa, and Naco formations. Because they are involved in part in structure affecting the alteration pattern, the red beds and associated brown sandstones have been examined more critically, although, even here, the Permian age assigned is inferred rather than proven

The relationships of the major igneous rock types have received more consideration, since knowledge of the sequence of formation of these rock units is essential to the work at hand. Although the dike rocks have been studied more carefully than the grouping adopted would imply, the sorting of the dike types and their separate mapping would constitute another geological problem of considerable magnitude.

Two zones of structural weakness appear in the Silver Bell Range. (1) A northwesterly trending belt of Paleozoic sediments produces an irregular escarpment on the northeast side of an irregularly depressed trough. On the southwest side the escarpment consists of a line of alaskite buttes and ridges called the "alaskite front" by Kingsbury, Entwistle, and Schmitt.

Along this front the alaskite is in many places close to the sedimentary formations, but only occasionally in contact with them. Usually, the separating rocks are belts of dacite porphyry or quartz monzonite believed to have been intruded along the zone of structural weakness forming a trough between two invaded rock masses. This zone was invaded first by dacite porphyry and later by quartz monzonite.

(2) An almost east-west structurally weak zone is noticeable at the southern end of the range on either side of the Oxide deposit. West of the Oxide deposit this zone forms a channel along Gifford Canyon with a limestone escarpment on the north and alaskite of Wild Hog Butte and Copper Butte on the south.

Quartz monzonite has been intruded along the second but more poorly defined trough in areas of varying size extending for more than 3 miles. The zone of weakness along which the monzonite has been intruded continues eastward from the Oxide deposit, but the alaskite exposures terminate at the deposit. To the east the Paleozoic sediments are also missing.

The most noteworthy feature of the two troughs is the belt of weakness provided for the invasion of the quartz monzonite and, of greater importance, the mineralizing solutions that followed. Although some mineralization evidently accompanied the introduction of the dacite porphyry and early dikes, the main epoch connected with the disseminated deposits was post monzonite.

The structure of the two zones of weakness

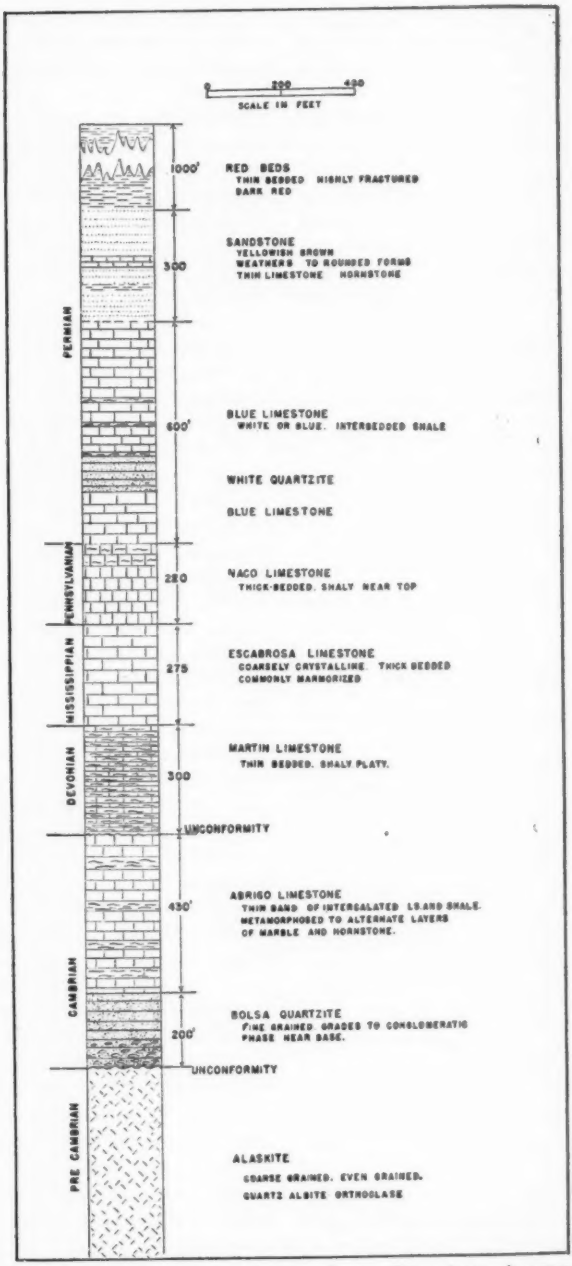

FIGURE 2.—STRATIGRAPHIC COLUMN, SILVER BELL AREA, ARIZONA

has not been completely determined, but both seemingly represent belts of faulting. Several generations of faulting are indicated: (1) when the alaskite was thrust against the Paleozoic sedimentary formations; (2) the dacite porphyry was intruded; (3) accompanying the intrusion of the quartz monzonite. In places, narrow bodies of quartz monzonite a few feet across follow faults between alaskite and dacite porphyry, or between alaskite and Paleozoic sedimentary rocks. Elsewhere the quartz monzonite areas have assumed sizable dimensions. (4) After the quartz monzonite was intruded, fractures parallel to the trend of the limestone escarpment formed channels for the invasion of late andesitic dikes (Pl. 3, fig. 1).

The dikes of the Oxide area are thought to reflect the structural pattern of thrusting to the south. Several more recent faults well exposed in the Oxide workings follow the same pattern (Fig. 3). While a single thrust plane may have existed at the first stages of deformation, the zone of principal deformation now appears to be made up of a somewhat obscure pattern of short interlaminated thrusts.

Another feature in the Oxide area is the cross fault along Copper Girl Canyon, with dacite to the east and alaskite to the west (Fig. 5). A minor parallel cross fault is shown in Figure 1 of Plate 2. South and west of the Oxide area a long fault of considerable displacement separates the dacite from the red beds.

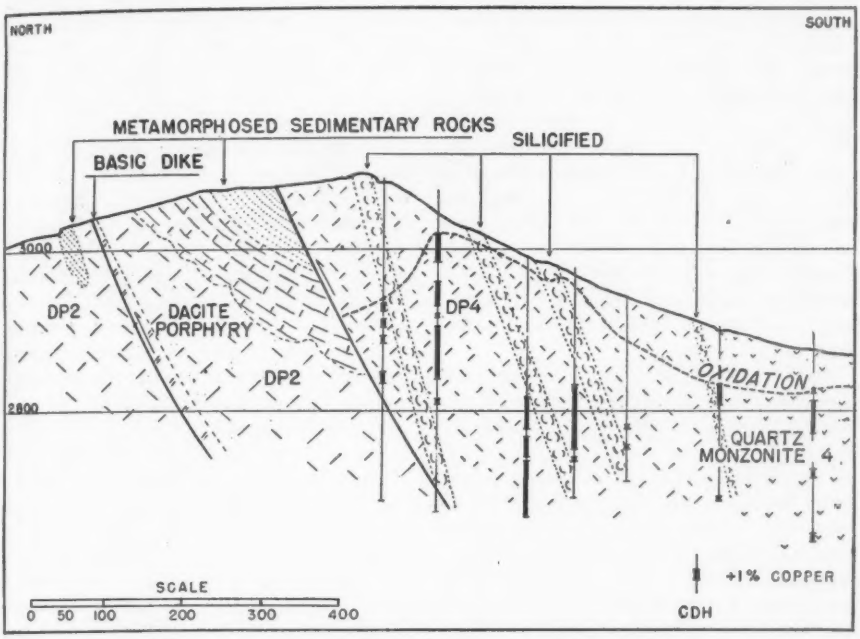

FIGURE 3.—SECTION THROUGH A PORTION OF THE OXIDE DEPOSIT

In the El Tiro area an en echelon group of faults occurs along the contact between the alaskite and the dacite. The lack of continuity of faults in the Silver Bell district received early recognition (Stewart, 1912, p. 483), but early workers apparently did not realize that the faults while often individually discontinuous may form a pattern as a group and that zones such as the "alaskite front" and the "Oxide trough" may extend for several miles.

Sedimentary Formations

General statement.—The best exposures of the Paleozoic sedimentary rocks in the vicinity oc-

cur in the Waterman Range to the south. On the south slope of the main canyon, southeast of Waterman Peak, the Bolsa quartzite rests on Precambrian granite. Many of these exposures are faulted, but in at least one outcrop, some 50 feet in length, the unconformity is well exposed.

Along the road for half a mile below the mine are excellent exposures of lower Paleozoic rocks. The northern margin of the Waterman Range also exposes brown sandstone and red beds overlying the Paleozoic limestones. Because of their stratigraphic position just above the limestone, and on account of the frequency with which the red beds in the southwest belong to the Permian, these strata are probably Permian. The beds have not been studied in detail by previous workers, although Kingsbury, Entwistle, and Schmitt (1941) referred to them as Cretaceous.

Bolsa quartzite (*Cambrian*).—This formation is estimated to be on the order of 100 to 200 feet thick. It forms prominent brown outcrops at the base of the Paleozoic section. It is fine-grained for the most part, but grades into a conglomerate phase near the base. Mr. Harold Courtright, after microscopic studies in connection with the churn drilling program at the Oxide deposit, points out that the zircon crystals in heavy-mineral concentrates from the Bolsa quartzite are considerably rounded.

It is not always certain whether the quartzite found in the vicinity of the Oxide and El Tiro deposits belongs to the Bolsa formation, or whether it represents the silicification of shaly limestone higher in the section. The quartzite is also readily confused with zones of silicification in the quartz monzonite and dacite porphyry.

The most prominent exposures of quartzite in the Silver Bell Range are found at El Tiro on the crests of Jesuit and Quartzite hills and near by. Good exposures also occur along Gifford Canyon north of the Oxide area.

Abrigo limestone (*Cambrian*).—The Abrigo limestone formation is reported by Kingsbury, Entwistle, and Schmitt (1941) to be about 430 feet thick. It is composed of quarter-inch bands of intercalated limestone and shale. Contact metamorphism frequently converts this to alternate layers of marble and hornstone. Differential weathering produces parallel grooves on the outcrops. The formation is best exposed along the west side of Gifford Canyon in the intermediate area between El Tiro and Oxide deposits.

Martin formation (*Upper Devonian*).—This is a thinly-bedded shaly limestone with some beds 6–12 inches thick. However, thin, shaly, laminated strata a quarter to half an inch thick predominate, so that the rock breaks into flat plates. The strata are easily eroded. The formation underlies a considerable portion of the depressed area between the alaskite ridges and the limestone escarpment in the intermediate zone between the two copper deposits. The maximum exposed thickness is about 300 feet.

Escabrosa limestone (*Mississippian*).—The Escabrosa formation comprises about 275 feet of coarsely crystalline thick-bedded limestone. Contact metamorphism has caused extensive marmorization of the rock. In places it is streaked gray or black, and in the Waterman Peak section a limestone, which is believed to correspond, is also gray to a considerable degree. Presumably the white color is an index of the extent to which contact metamorphism has affected the rocks of the sedimentary series. The formation occurs along the escarpment for about a mile northwest of the Oxide deposit.

Naco limestone (*Pennsylvanian*).—The Naco limestone is similar to the Escabrosa in the thickness of bedding, but it contains bands of shaly limestone varying in thickness and distribution. A 50-foot bed of shaly limestone occurs near the top of the formation. The formation is about 220 feet thick and is well exposed in the same region as the Escabrosa limestone.

Permian (*?*).—A sedimentary series consisting of a blue limestone at the base, about 100 feet of quartzite, and 450 feet of blue limestone,

PLATE 2.—EXPOSURES IN THE OXIDE AREA

FIGURE 1.—BARITE-BEARING FAULT PLANE ALONG A CONTACT
Between alaskite and quartz porphyry

FIGURE 2.—VIEW OF OXIDE AREA WITH COPPER GIRL HILL IN FOREGROUND

FIGURE 3.—ALTERATION VEINLETS ALONG FRACTURES CUTTING ALASKITE
On west slope of Mt. Hope

FIGURE 1

FIGURE 2

FIGURE 3

EXPOSURES IN THE OXIDE AREA

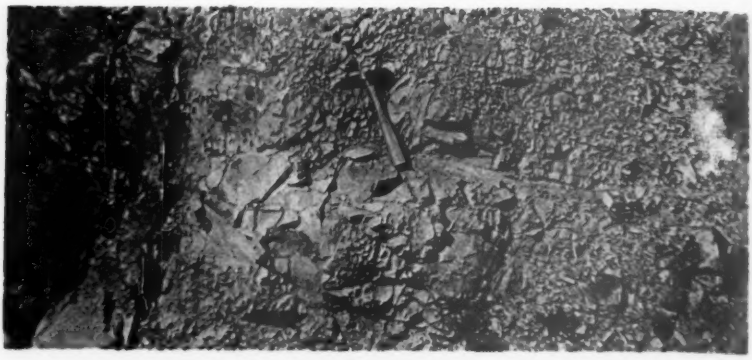

FIGURE 1

FIGURE 2

FIGURE 3

EXPOSURES IN OXIDE AREA AND EL TIRO AREA

shaly strata, and porcelainlike hornstone was tentatively assigned by Kingsbury, Entwistle, and Schmitt (1941) to the Permian. It overlies the Naco and is found along the limestone escarpment.

South and east of the Oxide area is a yellowish-brown sandstone in fault contact with the red beds but possibly underlying them. A complete section unbroken by faulting appears doubtful, and exposures yield no satisfactory group of attitudes. This series is estimated to be about 300 feet thick. A small amount of blue limestone accompanies the series, and along the canyon below the Oxide deposit porcelainlike strata may be found.

South of the brown sandstone is a section almost 2 miles in length along which red shale and sandstone are exposed particularly in the bottoms of dry washes. It has not been determined whether these strata contain fossils characteristic of the Permian, or silicified wood characteristic of the Cretaceous. The beds were penetrated in drilling for water at the Asarco camp site. The total thickness of the red bed series has not been determined, but probably it is considerably over 1000 feet.

Igneous Rocks

Alaskite.—The oldest and most prominently exposed (Pl. 2, fig. 2) igneous rock in the Silver Bell area is alaskite. Kingsbury, Entwistle, and Schmitt (1941) believed the alaskite to be Precambrian. While this appears the most reasonable interpretation, observed relationships are not so definite as the relationship between the granite prophyry and the Bolsa quartzite in the Waterman Range.

A granite porphyry similar to the granite porphyry of the Waterman Range is found in the Stump mine, 7 miles northeast of Silver Bell. Both types contain salmon-colored feldspar phenocrysts up to 2 inches long. In addition, the groundmass is filled with biotite.

In contrast to the granite porphyry, the normal alaskite is free from mica and the texture is even-grained. Only in limited areas is biotite prevalent, usually close to the quartz monzonite. Where monzonite is not observed on the surface the suspicion exists that it may underlie the alaskite. In one area, however, the alaskite is abnormally rich in biotite. In an adit in the El Tiro area, with a portal along the canyon west of Jesuit Hill, the alaskite resembles the rock from the Stump mine aside from the porphyritic character.

Faulting and metamorphism are so prevalent along the contact zone between the alaskite and the Paleozoic formations that normal sedimentary relations are obscured. Nowhere, however, is evidence definite to show that alaskite invaded the Paleozoic rocks. On the other hand, gouge, brecciation, and fracturing are common, particularly underground, which indicates that the alaskite was faulted up against or over the sediments.

Tolman (1909) sketched the geology of the contact-metamorphic ores of the Union, Billy, and Mammoth mines during early operations. At that time the alaskite was considered intrusive although the studies were limited.

Dacite, dacite porphyry, and early dikes.—The dacite is the oldest post-Paleozoic igneous rock. It has the texture and distribution of a flow rock and is believed to overlie the brown sandstone and the red beds. Probably it is Cretaceous.

Other flows overlie the dacite on the mountain northeast of the Oxide area. One is a widely exposed columnar lava which occupies the crest of the range.

Dacite is found for several miles along the south end of the Silver Bell Range. Exposures may be observed between the red beds and quartz monzonite along House Canyon south of Copper Butte. On Portland Hill and in the area of the Oxide deposit it has been invaded by both the quartz monzonite and the dacite porphyry. East of Portland Hill it occurs on either side of the intrusive zone.

The dacite porphyry invades the dacite and areas of Paleozoic sediments, and even the

PLATE 3.—EXPOSURES IN OXIDE AREA AND EL TIRO AREA

FIGURE 1.—LATE ANDESITE DIKE CUTTING QUARTZ MONZONITE West of Copper Butte

FIGURE 2.—QUARTZ VEIN, EAST END OF COPPER GIRL HILL

FIGURE 3.—ALTERATION ALONG FAULT PLANE IN QUARTZ MONZONITE PORPHYRY

alaskite in the marginal zones constituting the most widely exposed rock in the area. It is, in turn, cut by a swarm of dikes of intermediate composition and in belts of structural disturbance and alteration has been cut by the quartz monzonite. The quartz porphyry of Stewart (1912, p. 475) is probably the rock mapped later as dacite porphyry.

A most interesting pattern of dikes is superimposed on the dacite porphyry north of the Oxide area (Fig. 5). The trend is east-west, and the inclination is mainly southward. On the southern margin close to the mineralized belt the dikes blend into a pattern of faults suggesting intrusions of dike material along fractures sympathetic with the faulting on the south.

In the limestones along the margins of the dikes, and in the vicinity of the dacite porphyry, one finds considerable evidence of contact metamorphism. Tactite, hornfels, marble, and silicification are erratically but widely distributed. Some copper mineralization is associated with the tactite.

The early dikes are more widely distributed in the vicinity of the Oxide deposit than at El Tiro. At El Tiro the quartz monzonite is a prominent dike material. In the Oxide area however, the quartz monzonite is a comparatively uncommon dike material. The early dikes with their pronounced east-west trend cut the dacite, the dacite porphyry, and the limestone series and even, in some instances, the alaskite. They do not, however, appear to cut the quartz monzonite. In one place a short stub of an early dike appears along the border of a mass of quartz monzonite. This appears to have suffered assimilation. The early dikes show an unmistakable preference in distribution for the region east of the so-called "alaskite front."

Quartz monzonite.—Considerable areas of quartz monzonite are exposed in both the Oxide and El Tiro areas. Field relationships indicate that alaskite, Paleozoic sediments, dacite, dacite porphyry, and early dikes are all cut by intrusions of this rock. The rock is not entirely uniform throughout the area either in composition or texture. In the El Tiro area the quartz monzonite contains noticeably more coarse orthoclase than in the area to the southeast.

The quartz monzonite of recent workers in the Silver Bell area is probably the rock type mapped by Stewart (1912, p. 473) as biotite granite.

Dikes of quartz monzonite cut the limestone with pronounced metamorphism and mineralization in the vicinity of the Mammoth and Union mines of the El Tiro area. Near the mouth of the first canyon south of the Mammoth Canyon a dike of quartz monzonite porphyry cuts the quartz monzonite. Orthoclase crystals, presumably late, are prominent throughout the quartz monzonite of the El Tiro area. Late orthoclase is less apparent in the Oxide area except in thin sections.

It seems likely from the above relationships that the quartz monzonite of the El Tiro area represents a period of crystallization somewhat later than the quartz monzonite magma.

Stewart (1913) believed the rock mapped as quartz monzonite which he called biotite granite porphyry to be a differentiation product of the magma from which the alaskite came, but he recognized the intrusive relationship between the quartz monzonite and the alaskite. There is no record, however, that Stewart investigated the relationship between the alaskite and the Cambrian quartzite on the basis of which the alaskite is described as Precambrian.

Metallic Mineralization

The copper mineralization of both the Oxide and El Tiro deposits largely centers around chalcocite. This sulphide on polished surfaces may be observed replacing pyrite and chalcopyrite. Pyrite is by far the most abundant sulphide. Masses of it have been cut by veinlets of chalcocite and have also suffered "chemical brecciation." In concentrates, these minute lumps of pyrite, broken from a mass of chalcocite, frequently enclose an envelope of the copper sulphide. Pyrite also contains inclusions of chalcopyrite.

Occasional specks of bornite are visible in the concentrates, together with a few grains of an isotropic gray mineral, possibly tetrahedrite. Some covellite is also seen.

Below the enriched zone concentrates consist largely of pyrite and chalcopyrite. In such places the copper content falls considerably below any recognizable commercial grade. However, where chalcopyrite is abundant, the copper values frequently exceed 1 per cent.

Occasional small lumps and dendritic growths of native copper occur in the concentrates. Cuprite is also fairly common in the concentrates from the ore zone. Chrysocolla, malachite, and azurite are other oxidation products. Blocks of limestone and other Paleozoic sediments appear to have been caught and isolated in some places during the deformation.

In general, the limestone escarpment has been uptilted, and the strata inclined to the northeast away from the alaskite. Of course, numerous cross breaks occur, but this is the prevailing trend. The principal focus of such deformation as the district provides is brought to bear on the region to the west of the escarpment where the resultant effect has been the upthrust of the alaskite against the Paleozoic.

Numerous minor structures may be of considerable importance locally, but are not well enough exposed to enable recognition of any particular pattern.

The limestone strata are warped southeast of the Mammoth mine. Changes in attitudes indicate two zones of minor uplift extending northward from the end of the escarpment—one west of the Mammoth mine, the other east of the Union mine. The eastern uplift lies at the south end of the Union alteration channel, while the western uplift appears related to the El Tiro channel.

Relationship of Hydrothermal Alteration to Mineralization

Aside from the contact zones, in both the Oxide and El Tiro areas and to a lesser degree in the Union area, the belt of hydrothermal alteration includes the major indications of metallic mineralization. Such indications include data on copper content as shown by drilling, or the distribution of copper-bearing areas in mine workings and prospects. While assay records are privileged and are not available for reproduction the examination of a considerable number of assays points to at least general agreement between the zones of hydrothermal alteration and copper mineralization. This may be broadly substantiated in the field by noting the parallelism between the pattern of drilling and the limits of the more intense hydrothermal alteration. The examination of accessible mine workings, together with the inspection of prospects, provides further confirmation.

A considerable portion of the alteration belt on the Oxide area is shown in Figure 4. The belt is outlined, on the basis of advanced alteration, into four igneous rock types, Stages 3–4 in the quartz monzonite, 3 in the alaskite, 3–4 in the dacite, and 3–4 in the dacite porphyry. A few minor areas of lesser alteration are enveloped within the major zone, but these are too small to be considered in the major pattern. The belt extends for some distance east of the area mapped where it was followed in a reconnaissance. The darker area in Figure 4 indicates the area in which most of the drilling was concentrated.

A corresponding pattern for the El Tiro area is shown in Figure 7. The areas of more intense alteration form a discontinuous pattern in contrast to the continuous pattern of the Oxide area. Here the areas for which exploration data have been available lie between Mt. Hope and Mt. Expectation and in abandoned mine workings north of Mt. Hope. The isolated patches of alteration north of Jesuit Hill are unexplored.

The corresponding pattern for the Union area may be inferred from Figure 8. It would comprise the belt marked DP 3 and would include the transverse quartz monzonite dikes where alteration is in Stages 3–4. Scattered drill holes and a number of prospects are located in this area. On the whole the copper values in drill holes have been low, and the metallic mineralization observed in the prospects is chiefly pyrite.

Alteration

Effects in Igneous Rock Types

Alaskite.—The alaskite, a coarse-grained quartz-orthoclase-albite rock, is found along the southwestern slope for a large part of the length of the Silver Bell Range (Pl. 2, fig. 2). It has been classed as Precambrian, although the title to such antiquity is somewhat obscure. In any event it is the oldest igneous rock in the area, and it is probably older than the Paleozoic rocks. Further evidence on the age of this rather widespread and most important rock unit is to be desired.

The alaskite is intruded by dacite porphyry and quartz monzonite porphyry and is cut by at least two generations of dikes. It has been pushed upward against the Paleozoic sediments along a wide front. If once covered by Paleozoic rocks, as one may suppose, the old sediments have long since been eroded. At the northwest end of the range, faulted blocks of Paleozoic sediments are found in the alaskite. In places these are separated from the alaskite by a thin band of intrusive dacite porphyry or quartz monzonite porphyry. Some of these bands are little more than a foot thick and appear to occupy old faults.

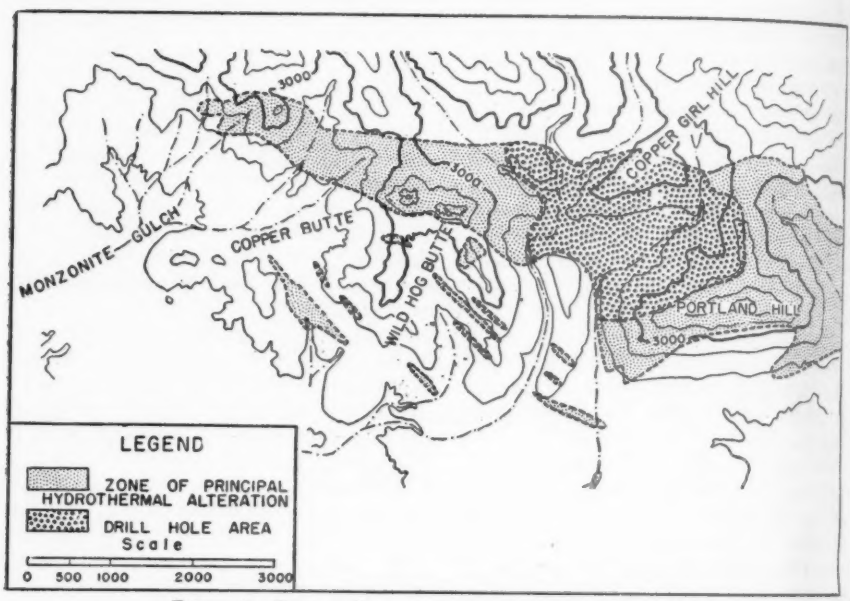

FIGURE 4.—BELT OF INTENSE ALTERATION IN THE OXIDE AREA

The alaskite is the most copper-stained rock in the area. It is also the most coarsely jointed. Likewise, it tends to resist erosion with the formation of prominent ridges and peaks capped by rock ledges. In some places, particularly at Copper Butte and Wild Hog Butte, fractures have formed channels for copper-bearing solutions. Where blocks have broken away, leaving exposed fracture surfaces, these are frequently coated a vivid malachite green. The coloration is probably out of all proportion to the copper content of the rock. On the other hand, where stained alaskite overlies highly altered quartz monzonite porphyry, the elements of a more significant copper mineralization exist at a lower level.

The most quickly and easily recognized mineral is light-gray quartz that stands out on weathered surfaces, unless the rock has been subjected to advanced hydrothermal alteration. The gray grains, about a quarter of an inch across, are also clearly visible on freshly broken surfaces. The quartz grains are ordinarily more abundant and coarser in the alaskite than in the quartz monzonite porphyry and the dacite porphyry.

Orthoclase is light gray or almost colorless in unweathered and unaltered specimens, but ordinarily as found in the outcrop is stained dark reddish brown, or at least a salmon pink. It is even coarser than the quartz. Specimens collected retain reflectivity along cleavages, often even when stained by weathering. The mineral is less resistant to hydrothermal alteration than the quartz, and more resistant than the albite.

The plagioclase feldspar is albite as determined optically and by comparison of X-ray

diffraction patterns with known albite. It varies from white to pale cream color on freshly broken surfaces and is usually cloudy. Albite is the first major mineral in the rock to be affected by hydrothermal alteration and may be completely replaced while much of the orthoclase is still unaffected.

Probably the least altered alaskite occurs on the ridge southwest of Copper Butte (Pl. 2, fig. 2) and in the intermediate region between the Oxide and El Tiro areas. Even specimens collected in the field as unaltered, when examined with the microscope, show traces of alteration effects. These are most evident in the albite, which contains traces of sericite and hydromica. Fractures are also apt to be filled with small veinlets of alunite and jarosite. Presumably the alunite is due to the breakdown of the orthoclase at lower levels within the rock mass and, at the same time, the introduction of sulphate-bearing solutions. The two combine to deposit alunite at a higher level and in essentially unaltered rock. In areas of more advanced alteration the alunite and jarosite may form small veins up to an inch thick. Another fracture-filling mineral indicative of more advanced alteration is barite. Highly altered areas of both alaskite and quartz monzonite porphyry may contain veins of barite as much as 3 inches thick. Small cleavages of barite may be found in relatively unaltered areas of alaskite. In view of the placement of the vein barite in the alaskite and quartz monzonite and its close association with the alteration, it is believed to be hypogene.

Three stages of alteration are recognizable in the field. Alaskite (1) is essentially free from alteration. Alaskite (2) has suffered some hydrothermal alteration with destruction of the albite. Alaskite (3) has suffered the alteration and destruction of both the albite and the orthoclase. In limited areas the entire rock mass has been altered with the introduction of sericite, quartz, and some pyrite. Since a corresponding advanced alteration occurs in each important intrusive rock, this stage is mapped where observable as a quartz-sericite zone. This phase is probably more extensive than the map indicates. In the mine workings or in diamond-drill cores it would be more easily recognizable.

The most common alteration minerals in the alaskite are sericite, hydromica, and alunite. Jarosite is fairly common. Barite is irregularly distributed as a vein mineral. Opal and halloysite (?) are found in veinlets. Limonite after pyrite, and leucoxene after sphene are frequently observed.

Three accessories—sphene, zircon, and apatite—are seen in the fresh rock. Zircon and apatite persist even in the alteration of Stage 3.

The most highly altered alaskite occurs in the vicinity of highly altered quartz monzonite porphyry. Advanced alteration of the alaskite is uncommon when the adjacent rock is fresh or slightly altered quartz monzonite porphyry.

Alteration banding is prominent in the El Tiro area, particularly on the west slope of Mt. Hope (Pl. 2, fig. 3). Quartz veinlets a fraction of an inch thick and from a few inches to several feet in length form a striated structure with a strike of N. 80° E. and dip of 80° S. Intermediate between the quartz veins are layers of alaskite 2 and alaskite 3. Near highly altered quartz monzonite prophyry, however, the entire rock is apt to be alaskite 3. Alunite, barite, and occasional streaks of limonite are fairly common.

Biotite occurs in the alaskite in some places. It has been observed in the vicinity of comparatively unaltered quartz monzonite porphyry west of Copper Butte and west of the El Tiro area. The proximity of the biotite to the monzonite contacts and the absence of the mineral in areas more remote from the contact suggest that the biotite is a contact effect related to the introduction of the monzonite. Thus, in areas where monzonite is not exposed, or may be covered by mantle rock, the presence of biotite may indicate the presence of near-by monzonite. Where the alaskite is highly altered the biotite appears to be similarly altered. Thus, in contact areas, this mineral where found in the alaskite may furnish an index to the degree of alteration of near-by or buried monzonites.

In alaskite 3, a large amount of the quartz retains the texture of the original alaskite. The feldspars, on the other hand, are entirely altered to a claylike mass. In the alaskite 2 only the plagioclase is claylike.

Frequently small aplite dikes cut the alaskite. These do not appear to affect the extent of mineralization in the rock. Pegmatites are conspicuously absent. Apparently all traces of copper in the rock and evidences of hydrothermal mineralization have been inherited from

subsequent activity due to the after-effects of later magmas. From the standpoint of hydrothermal after-effects due to the alaskite magma, the rock is believed to be essentially dormant. In places, however, it may be a good host for the reception of the after-effects of the monzonite invasion.

Dacite.—The dacite is a dense, almost glassy, flow rock containing small scattered phenocrysts of pinkish feldspar. It varies from dark gray to almost black.

It is exposed south and east of the Oxide area both unaltered and in various stages of alteration. The rock has been invaded by dacite porphyry and quartz monzonite. It lies in contact with the alaskite, but, in each instance, the actual contact line is believed to be a fault. Sedimentary rocks are found south of the dacite exposures. Since these rocks continue without interruption to the Paleozoic exposures of the Waterman Range, where they rest upon Precambrian, it seems natural to place the volcanic rock above the entire sedimentary series. The actual age, however, is not definitely fixed, although it may be Cretaceous. It is older than the quartz monzonite which is believed to be typical of the Tertiary intrusives of the region. The dacite porphyry which invades the dacite may also be Cretaceous. Dacite has not been observed in the area mapped at El Tiro, although a similar lava may be observed to the north.

The dacite alters first to a fine greenish rock that still retains the brittle character of the original lava. Next, it appears to be more definitely chloritic and no longer brittle. In each case, the rock is greenish, and close scrutiny reveals relict outlines of small dispersed original feldspar phenocrysts. Both stages are grouped as dacite Stage 2.

In final alteration the greenish color largely disappears, and fresh exposures are white and claylike. In the field, this stage is not distinguished with assurance from the final stages of alteration of the dacite porphyry. Both occur in the Oxide area in contact with quartz monzonite Stage 4. In general, the dacite occupies the area in the vicinity of Portland Hill, while the dacite porphyry is found on Copper Girl Hill and northward.

Copper mineralization has been found in both the dacite and the dacite porphyry in the highest stages of alteration. However, the rock is not considered as favorable a host for the higher grades of mineralization as is the quartz monzonite Stage 4. The occurrence of the better copper values in the holes penetrating the quartz monzonite Stage 4 on the north side of the Oxide deposit, rather than in the dacite to the south, supports this viewpoint.

The alteration stages in the dacite may be summarized as follows:

Stage	1	Aphanitic and unaltered
Stage	2	Chloritized
Stage	3-4	Largely argillic (3) sericitic and silicified (4)

Dacite porphyry.—The dacite porphyry widely exposed on the north side and along the northern margin of the alteration zone in the Oxide area, also occurs extensively in the El Tiro area, where the principal exposures are east and north of the belt which includes the Hilda, Lauzanne, Williams, Daisy, and Kurtz mines.

In Hilda canyon a short distance below the Yankee shaft of the El Tiro area inclusions of earlier alaskite occur in the later dacite porphyry.

In the Oxide area quartz monzonite occurs along the alteration zone and to the south. Although the quartz monzonite porphyry is the younger rock and, in a few instances, invades the dacite, it is not found north of the alteration zone. Similarly, with certain minor exceptions, the dacite porphyry is conspicuously absent on the south. Prior deformation may have displaced the dacite on the south before the monzonite invasion.

Large areas of the dacite porphyry are unaltered except for the rather widespread distribution of epidote. Dacite areas, however, have been widely invaded by a system of early dikes that presumably represent the end stages of the dacite magma and appear to have formed along fractures which, in turn, are sympathetic with earlier deformation. In fact, the unusual distribution of the dike pattern in itself is thought to have structural significance. A considerable amount of contact metamorphism in the limestone may be attributed to the action of the early dikes. Much of the epidote, which is so broadly distributed in the dacite porphyry, may have been formed at the same time.

The invasion of the dacite porphyry may have been preceded by an outflowing of dacite lava. Such lava is found in considerable quantity both north and south of the alteration zone, particularly east of the Oxide deposit. South of Portland Hill, the dacite lava appears to rest upon sedimentary rocks which have been generally classed as Cretaceous. North of the alteration zone, the dacite porphyry appears to invade the dacite lava, as indicated by the exposures on the south slope of Hook Dike Mountain.

A summary of the field relationships would thus place the invasion of the dacite porphyry after the dacite lava (or dacite) and before the system of early dikes. It is found in intrusive contact with the alaskite, particularly in the El Tiro area, and cuts the Paleozoic sedimentary formations in a number of places. Possibly it is Cretaceous in age.

Hydrothermal alteration of the dacite porphyry is apparently attributable to the post-monzonite hydrothermal solutions so prevalent along the alteration zone. This is indicated by the proximity of the highly altered monzonite where alteration is advanced in the dacite porphyry. Marginal features along contacts, and the distribution pattern are likewise significant.

A number of old drill holes in the Oxide area have been drilled in the most highly altered stages of the dacite porphyry. These indicate copper mineralization, in some instances as great as may be found in the quartz monzonite porphyry. The distribution of the old workings in the El Tiro area also indicates that in places active mining may once have been carried on in the most highly altered dacite porphyry.

In general, even the most highly altered dacite porphyry has probably not provided as good a host rock for the deposition of copper mineralization as the quartz monzonite porphyry. This may be due to the finer texture and lesser permeability or to the dip of the alteration channels to the southward, away from the dacite porphyry mass.

While the dacite porphyry may be less prolific in copper mineralization, some productive areas indicate that the most highly altered phases of this intrusive should receive close study. Likewise, this unit should be kept in mind from the standpoint of exploration.

One of the most distinctive features of the dacite porphyry is the prevalence of small, rounded quartz phenocrysts, usually slightly less than a millimeter in diameter. In many cases the grains have a rounding or a subangularity of a type frequently associated with sedimentary rocks.

The feldspars are fine-grained and consist of orthoclase and less abundant albite. Smooth cleavage surfaces both striated and unstriated may be observed in hand specimens of the least altered dacite porphyry. Occasionally a few specks of minute platy biotite may be seen. Traces of epidote are nearly always present. The rock is a distinctive bluish gray and grades to a white or bleached appearance during alteration.

In the dacite porphyry, four stages of alteration are recognized in field mapping. The least altered gray dacite porphyry constitutes Stage 1. Bleaching of the groundmass, with removal of epidote and deposition of secondary chlorite, indicates Stage 2. Destruction of the chlorite by further bleaching, accompanied by a general whitening of the groundmass, occurs in Stage 3. The original quartz phenocrysts have largely, if not entirely, disappeared, and the groundmass is to a considerable degree argillic in Stage 4.

Another significant feature of the dacite porphyry is the presence of inclusions of various sedimentary rock types; the most common is quartzite. Limestone, if present, has been altered to epidote. Inclusions up to an inch in diameter are reasonably abundant.

The accessory minerals of the dacite porphyry as indicated under the microscope are ilmenite, rutile, apatite, and leucoxene. Leucoxene is formed by alteration of the other titanium-bearing minerals. Zircon is also present. Both rutile and zircon may persist even in the most altered dacite porphyry.

The clay mineral produced by alteration of the dacite porphyry has indices of refraction which correspond to hydromica. X-ray diffraction patterns and thermal curves confirm its identity. Small crystals of freshly crystallized orthoclase are found in the most altered dacite porphyry along with quartz believed to be secondary. Thus, apparently not only quartz monzonite porphyry and alaskite, but dacite porphyry as well, contain orthoclase (or adularia ?) of late hydrothermal origin.

Quartz monzonite porphyry.—The quartz

monzonite porphyry is the youngest of the principal intrusive rock masses in the Silver Bell Range. It has been cut by minor invasions of andesitic dikes, late rhyolite, and, in a single instance, a dike similar to the monzonite itself, but more highly porphyritic. The aftermath of the quartz monzonite porphyry invasion is believed to have provided the hydrothermal alteration which has altered most of the rock types in the area, and has also been responsible for the introduction of the principal copper mineralization.

The most distinctive mineral in the rock is lustrous black biotite. It often forms small books about 0.1 inch in width and of approximately equal thickness. Occasionally, however, the biotite is more platy, being composed of crystals twice as wide and about half as thick. Traces of alteration to chlorite are nearly always present, even in the least altered rock.

Biotite is one of the most significant minerals subject to alteration in the area (Fig. 9). Recognizable stages in the alteration of this mineral are: (1) ordinary black biotite; (2) biotite altered to green chlorite; (3) biotite altered to light-brown mica; and (4) biotite altered to sericite, or showing only relict outlines. In the ore-bearing stages, chalcocite occasionally fills relicts after biotite.

The progressive stages in biotite alteration provide indices readily identifiable for field mapping. Stages 1 to 4 in the alteration sequence of the monzonite are based on the biotite. Parallel alteration of the other minerals in the rock may be observed, but the stages of the biotite are most easily recognized in the field.

In addition to biotite, the chief minerals of the rock are quartz and oligoclase. At times orthoclase is also a prominent constituent. The latter is abundant in the El Tiro area and is found occasionally in thin sections of the quartz monzonite porphyry from the Oxide area. It may form coarse phenocrysts. In a single porphyry dike where orthoclase phenocrysts are prominent, the mineral is poikilitic and forms euhedral crystals an inch or longer.

Coarser quartz occurs in scattered grains up to 0.1 inch in diameter. Fine grains are also dispersed through the groundmass. The phenocrysts are much less abundant than the quartz in the alaskite, and the crystals are smaller.

The plagioclase in the rock is probably both oligoclase and andesine. It is zoned and tends to alter selectively, the more calcic phases altering first. Only cleavage surfaces are observable in the freshly broken unaltered rock. At stage 3 or 4, however, clay pseudomorphs after plagioclase may be broken from the original rock. These show small but clean-cut outlines of plagioclase crystals.

The accessory minerals in the rock are sphene, zircon, and apatite. Sphene forms small honey-yellow crystals in the unaltered rock. Zircon and apatite have only been observed with the microscope.

Considerable areas of quartz monzonite porphyry appear in both the Oxide and the El Tiro areas. Monzonite 2 is probably the most widely exposed of the monzonite stages of alteration, but Stage 4 appears to be by far the most important from the standpoint of copper mineralization. This stage appears to be either the host rock or a near neighbor wherever sufficient copper occurs to be considered as ore. Stage 4 monzonite forms the backbone of the alteration zone. Stage 3 appears to have a similar, but less effective, relationship. It is also much less abundant.

In Stage 2, chlorite is formed as a replacement of biotite, but in addition there are many veinlets of chlorite which cut the rock in all directions. Pseudomorphs of chlorite after biotite are common and form a distinctive feature. Chlorite makes the rock greenish gray in outcrops.

Calcite is fairly abundant because of the partial breakdown of the plagioclase. It accounts for a leached white coating on some of the outcrops. Under the microscope veinlets of the carbonate are seen cutting quartz and sphene in addition to feldspar.

At this stage the plagioclase has not been greatly destroyed. However, under the microscope, traces of kaolinite, hydromica, and sericite may be observed to form at the expense of the plagioclase. In the hand specimen the feldspars are frequently cloudy but are not claylike.

Stage 2 may change abruptly at a contact to either 1 or 3, even within a few inches. On the other hand, wide areas of Stage 2 show little or no indication of more advanced alteration.

Traces of copper mineralization are occasionally found, but pyrite is more common.

The accessories—sphene, zircon, and apatite—are not destroyed during this stage of alteration, but leucoxene is more abundant than in Stage 1.

In the field, Stage 3 monzonite is recognized through the observation of two principal criteria with the hand lens: (1) the biotite retains its original form, but has altered to a brownish mica with a color distinctive from the limonitic stain ordinarily associated with weathering; (2) the feldspars have lost luster and look distinctly claylike.

Outcrops of this monzonite frequently show a distinctive color contrast when compared with outcrops of Stage 2. On the other hand, the color distinction between Stages 3 and 4 is hardly apparent. A number of drill holes in the Oxide area indicate that this stage of alteration of the quartz monzonite porphyry, at least in certain instances, may be associated with ore deposition. The same would also appear to apply to El Tiro. This alteration stage is not so widespread as either the preceding or the following stage.

Under the microscope it may be observed that hydromica has been formed at the expense of albite. Kaolinite accompanies the hydromica, however. The coarse biotite crystals form hydromica, which appears white in the microscopic field. Sericite forms veinlets and replaces biotite. The principal replacement of biotite, however, is by hydromica. An interesting aspect of this stage is the formation of secondary orthoclase (or adularia ?) and quartz. The original quartz phenocrysts common to Stages 1 and 2 are largely destroyed in Stage 3. Apparently zircon is the only accessory mineral to escape alteration.

This stage is recognized in the field most readily by the complete transition in the areas which were originally biotite. The areas occupied by this mineral are no longer black, as in Stage 1, green as in Stage 2, or yellowish brown, as in Stage 3, but have turned white. In some instances even the white mica has been destroyed, leaving sericitic shreds along the borders of casts. These casts are occasionally filled with chalcocite in copper-bearing zones.

A fifth stage exists, but has not been separated from Stage 4. In this stage, silicification and sericitization have completely obliterated considerable areas of the original rock. Large areas of bedrock exposures would be required for mapping with any degree of accuracy. The stage should be mentioned, however, since, as exploration progresses, information regarding this stage will be more and more desired. This phase represents an apparent regression in copper content as contrasted with Stage 3 and the early part of Stage 4, when copper precipitation apparently was heaviest.

Oxide Area

General features.—The geology and alteration pattern of a portion of the Oxide area are shown in Figures 5 and 6, which represent selected portions of the alteration belt shown in Figure 4. The various stages of hydrothermal alteration in the alaskite, dacite, dacite porphyry, and quartz monzonite as well as areas of silicification are shown in the area which surrounds Copper Girl Hill.

On the north is an area broken by faulting and cut by numerous early dikes which contains remnants of Paleozoic sediments. Several thrust faults are shown even within the small area encompassed on the map. These typify a pattern which extends for considerable distances on either side. In addition to the well-marked faults mappable on the basis of fracture zones, gouge, and displacements of sedimentary strata, numerous earlier dikes appear to have occupied previously existing fault lines.

The area well within the zone of major faulting is distinguished by absence of the effects of hydrothermal alteration. In this region the dacite porphyry is well crystallized with feldspars which exhibit brighly reflecting cleavage surfaces free from the effects of argillic and sericitic alteration. Here the chief alteration effect is of contact metamorphism in the carbonate sediments. Masses of garnet and epidote intermixed with quartz and secondary carbonate have been mapped as tactite. Apparently the effects of metamorphism responsible for the tactite are related to the intrusion of the early dikes and perhaps in some instances to the dacite prophyry. For the most part the tactite is distributed among the margins of the early dikes or in the vicinity of faults and fractures traceable to the dikes.

The thrust faults are largely inclined as illustrated in Figure 3 which represents a somewhat generalized cross section extending from the hundred such fractures bearing sulphides were counted and mapped in about 400 feet along a cross cut at the eastern end of Copper Girl

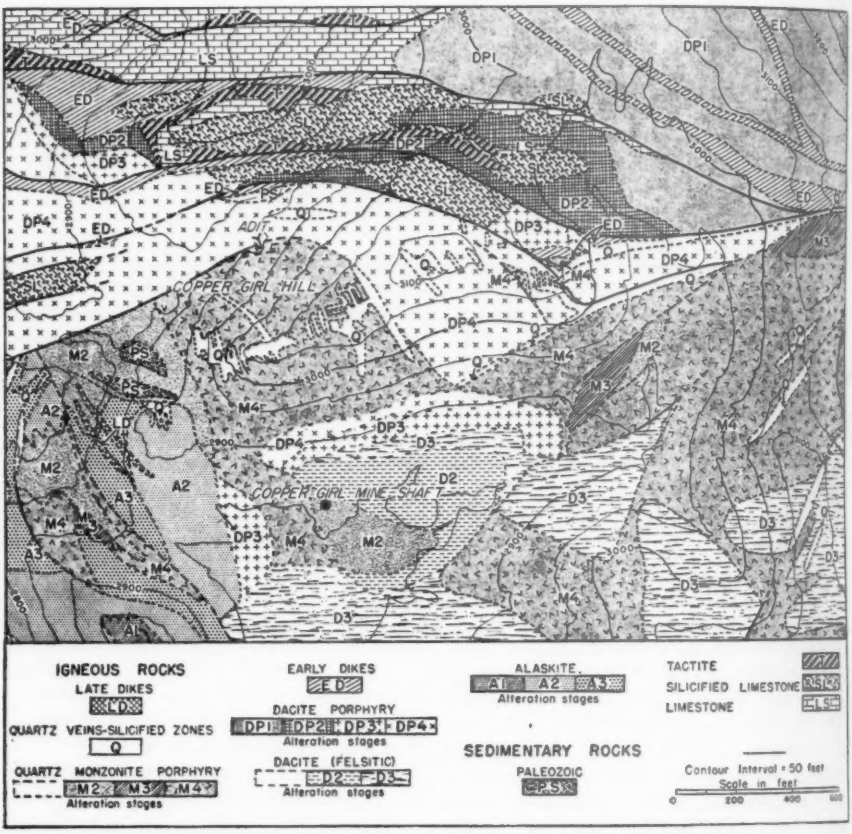

FIGURE 5.—STAGES OF ALTERATION IN THE VICINITY OF COPPER GIRL HILL

vicinity of the adit on Copper Girl Hill southeasterly toward the Gopper Girl mine shaft. Faults may be observed in a number of places. In one place an inclined shaft follows a fault line for some 200 feet. The dip of the faults at 50° S. or less is confirmed by the intersection of the fault planes with the topography as observed in surface mapping.

A number of smaller displacements may also be observed which are not indicated in Figure 3. Also minor fractures form a complementary pattern. Many of these have been filled with chalcocite, chalcopyrite, and pyrite. Several mine. The total sulphide content of the fractures measured would appear to represent about a third of the copper content of the section studied. The major portion of the copper is present in disseminated specks and minute seams scattered through the rock between the fractures.

The area containing the principal hydrothermal metamorphism as shown in Figure 5 includes a zone which cuts across Copper Girl Hill. The area designated by M3, M4, or DP4 represents the highest degree of hydrothermal metamorphism (Fig. 4). Certain areas, indicated

by Q, are almost solid quartz and are believed to constitute the final stage of major hydrothermal activity. The silicified zones are deficient in sulphide mineralization where observable in underground workings. The assays yield a lower copper content than noted in zones of argillic alteration on either side.

The quartz monzonite porphyry has invaded the zone of contact between the dacite porphyry and the alaskite west of the area mapped. To the east, however, the quartz monzonite invasion has ruptured a zone of contact between the dacite and the dacite porphyry. East of the Oxide area the same line of weakness along which the quartz monzonite has been intruded is bordered on both sides by dacite.

The inclination of dikes and faults and the fracture pattern of the alaskite indicate that the monzonite was probably not emplaced vertically but forced its way upward from the south and southwest. Hydrothermal alteration following the monzonite tended to rise in the same direction. As a result the overlying alaskite has been highly altered in the immediate vicinity of contacts with the monzonite of Stage 4.

A large mass of least-altered quartz monzonite is exposed in the southwestern part of the area (Pl. 1). Chloritization of the biotite in the rock and the distribution of chlorite in fractures is quite widespread. Traces of copper mineralization may be found in numerous places. Whether this rock mass becomes less altered or more altered at depth provides a worthy subject for consideration. The absence of isolated patches of intense alteration is largely an unfavorable factor in terms of metallic mineralization. While exploration of this area is unlikely now or for some time to

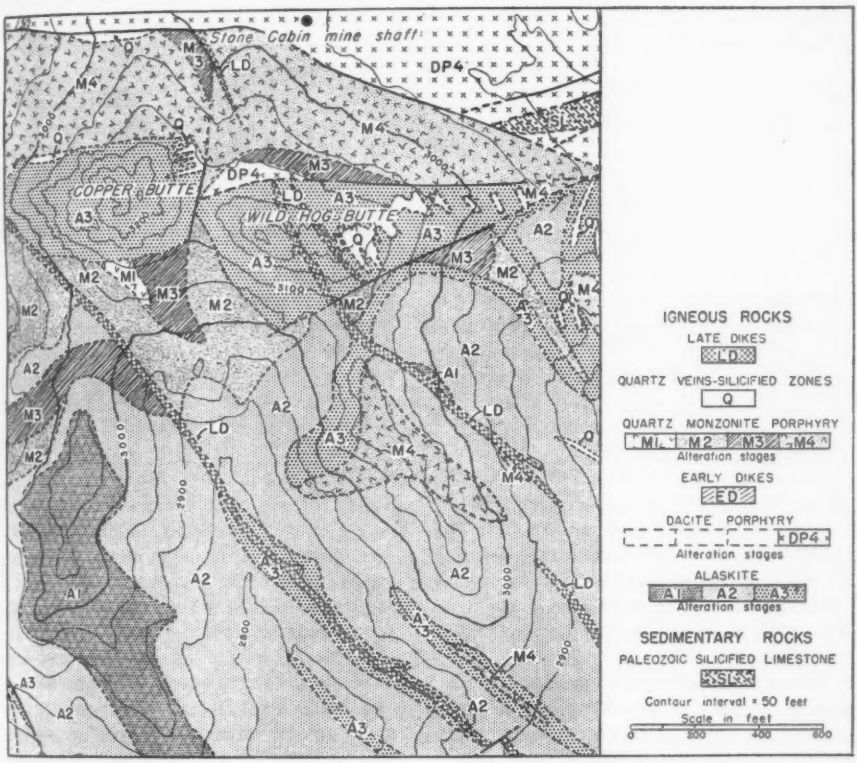

FIGURE 6.—STAGES OF ALTERATION IN THE VICINITY OF WILD HOG BUTTE

come, the situation should be kept in mind as exploration of the area progresses.

A considerable discordance exists between opposite walls of Copper Girl Canyon (Pl. 1). Alaskite lies to the west, while quartz monzonite, dacite, and perhaps dacite porphyry provide the chief exposures to the east. The principal difference, however, is that both the quartz monzonite and dacite on the east show a preponderance of high-intensity alteration, whereas the alaskite across the canyon is much less altered.

While a few small areas of quartz monzonite, Stage 4, are found in the west block, the rocks exposed are largely alaskite Stage 2. On the other hand, old churn drill holes record copper mineralization at the deepest levels in the Oxide area within the limits of this block. Data are inadequate for proof, but they suggest that faulting contemporaneous with the monzonite intrusion may have resulted in relative movement which placed the two blocks opposite each other in their present positions. The alaskite was uplifted, and the monzonite area down-dropped.

A belt marked by strong silicification and occasional quartz masses or veins (Pl. 3, fig. 2) extends along the crest of Copper Girl Hill (Pl. 2, fig. 2) and westward as far as North Butte. Likewise prominent fissure fillings of quartz occur on Portland Hill, southwest of Portland Hill, and in the area east of the map area. A prominent feature of this eastern area is a large quartz vein.

Belt of mineralization.—Previous mapping, correlated with the current study, indicates that the principal rock types exposed in the Copper Girl workings are quartz monzonite dacite porphyry, and dacite, all highly altered. On the other hand, the recognition of the alteration units in the old workings is accomplished with much less facility than on the surface.

In many instances in the Stage 4 quartz monzonite the relict forms after biotite have been replaced by chalcocite and, where present, are not so easily discernible as in leached outcrops on the surface. In addition, certain areas have been so silicified that even traces of the original textures have been destroyed. The dacite and dacite porphyry are recognized chiefly through contrast in rock texture and fracturing. The dacite rocks are easily fractured into small fragments, and the texture is almost dense.

Chalcocite is the predominant copper sulphide mineral in the old workings. Chalcopyrite is found on polished surfaces but is not noticeable in hand specimens. Molybdenite may be observed in a number of places, although assays indicate that the average molybdenum content is too low to be significant.

Unverified commercial reports indicate that mercury is present in minute amounts, but in this study no definite mineral source of mercury was detected. However, no particular deviation

TABLE 1.—COMPARATIVE COPPER CONTENT OF PYRITE-QUARTZ AND STRONG CHALCOCITE, ALONG NUMBERS 1 AND 2 CROSS CUTS, COPPER GIRL MINE

Location Distance from drift at incline	Number	Assay Per cent Cu	MoS_2	Type
10	134A	2.30	Trace	Disseminated Chalcocite
20	134B	.70	—	Disseminated Chalcocite
30	134C	.82	0.09	Disseminated Chalcocite
70	133A	0.28	—	Pyritic Zone
90	133B	0.16	Trace	Pyritic Zone
100	133C	0.35	0.05	Pyritic Zone
115	133D	0.29	—	Pyritic Zone
215	135	0.42	0.06	Pyritic Zone
280	136	0.98	0.05	Disseminated Chalcocite
315	137	0.15	Trace	Pyritic Zone
385	138	1.82	0.07	Disseminated Chalcocite
Adit Level	132	1.12	Trace	Disseminated Chalcocite

from the objective of alteration study was attempted for the purpose of contributing to the solution of the mercury problem.

Pyrite is much more abundant in some parts of the workings than in others. Likewise, where the pyrite is more abundant, the amount of chalcocite diminishes. Late quartz and silicification are also more prominent where pyrite is more abundant. The contrast between the mineralization in the pyritized areas and in the chalcocite-rich areas permits a fair separation of specimens from the two zones with a hand lens.

To verify this distinction 11 specimens from various parts of Numbers 1 and 2 cross cuts were submitted for assay (Table 1). In collecting the specimens every effort was made to eliminate visible veinlets of chalcocite. In this way it is believed that the type represented by the specimens would be clearly disseminated. This distinction is made because numerous chalcocite-filled fractures may be seen along the cross cuts, and it was considered desirable to confine comparison to mineralization of the disseminated type. Otherwise lump specimens would have little significance.

Stewart (1912, p. 498) calls attention to the replacement of pyrite by chalcocite at El Tiro with the formation of pseudomorphs of chalcocite after pyrite.

Although a way of confirming this has not been found, such quartz veining and silicification as are exposed on the surface above, with intermediate zones of Stage 4 monzonite alteration between, may correspond to the chalcocite and pyritic zones observed in the Copper Girl Mine. At least, this possibility should be considered in connection with the exploration of the Oxide area.

A cross section (Fig. 3) of a portion of the workings illustrates this interpretation. In common with other interpretations, this hypothesis must bridge a number of gaps for which information is not available. It is based upon observation along the cross cuts in the Copper Girl mine, a review of the old assays, assays of the two types of mineralization exposed, data from near-by churn drill holes, and surface exposures.

If this interpretation is correct, the Oxide deposit would consist of a series of steeply inclined and intercalated lean and rich zones of variable width. Pyrite would be the predominant sulphide in the lean zones, and chalcocite in the rich zones. The zones would appear to vary in width, with dimensions on the order of 20 to 50 feet. The dip would vary from 50° to vertical. The depth to which the chalcocite extends in these zones is as yet undetermined. Copper mineralization in significant amounts has been found throughout between the two extremes constituting a vertical range of over 300 feet in the Oxide deposit. It is not known whether this mineralization is predominantly chalcocitic or whether chalcopyrite is also prominent in places. In recent drilling the chalcocite zones have given way abruptly to chalcopyrite and pyrite at the bottom of each hole.

Chalcocite veinlets, varying in thickness from a small fraction of an inch to 2 inches, occur along the cross cuts in the Copper Girl mine. These suggest that most of the copper mineralization in the mine may be attributed to fractures filled with chalcocite. Further study indicates that this is probably not so. To estimate the significance of the easily visible fractures as a source of copper, the thicknesses of those along cross cuts 1 and 2 were plotted. The combined thickness of all veinlets one-tenth of an inch or thicker, for more than 400 feet of cross cut, would furnish not more than one-third of the copper estimated for the entire section. Microscopic study indicates that a certain amount of copper occurs as chalcopyrite included in grains of pyrite.

El Tiro Area

Belt west of Silver Bell.—Plate 4 illustrates the stages of alteration in a portion of the El Tiro area. The area shown is part of the region covered by the El Tiro alteration diagram in Figure 7 and embraces the terrain in the vicinity of the Daisy, Williams, Lauzanne, and Hilda mines. The Atlas mine area lies to the northwest, and the Mammoth and Union mines lie beyond the limits of the map to the east.

The map indicates the distribution of the various alteration units in the vicinity of the principal old workings in which the ores were largely of the noncontact metamorphic type. Since most of the old workings of the Mammoth mine, judging from the surface geology

and the large mine dump, were of the contact-metamorphic type, and the Mammoth mine lies to one side of the El Tiro chain of old workings, it is not included. The easternmost workings of the Union mine were of the disseminated type as indicated by the lower dump. The more extensive and western workings were in the contact zones.

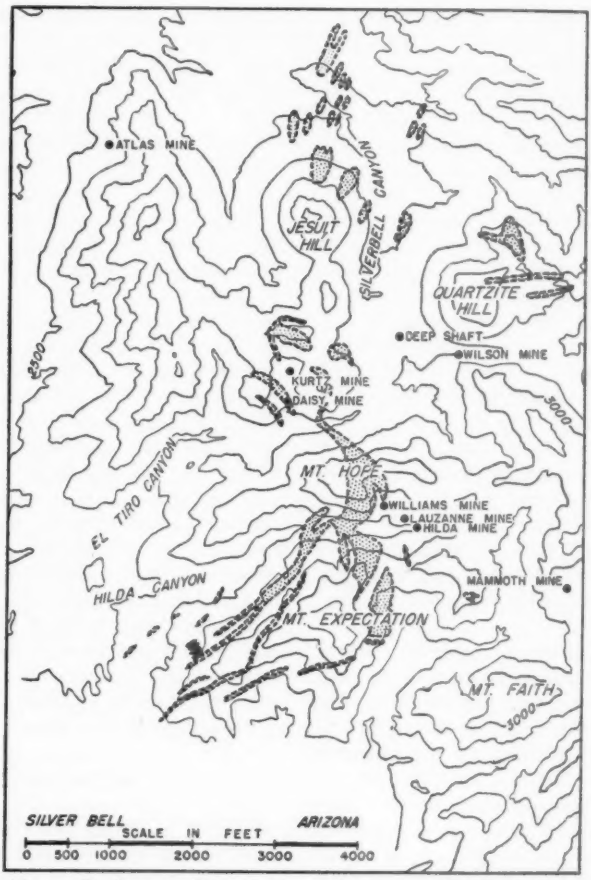

FIGURE 7.—BELT OF INTENSE ALTERATION IN THE EL TIRO AREA

In general, the alteration is much less intense in the El Tiro area than in the Oxide area as shown in Figure 4. The amount of quartz monzonite Stage 4 exposed is small. The proportionate amount of lower-stage alteration in the monzonite exposed is larger. On the other hand, copper mineralization appears to be found in rocks whose grade of alteration is lower than in the Oxide area. Further, broad zones of brecciation and near-by fissure zones, as shown by the old workings, were quite extensively mineralized.

In the El Tiro area, as in the Oxide area, the loci of greatest alteration are directly associated with the more intense copper mineralization. The surface map of the alteration zones is believed to be a fair index of the most likely area to yield ore underground.

The significant altered rocks appear to be for the most part quartz monzonite Stages 4 and 3;

dacite porphyry Stages 4 and 3; and alaskite Stage 3. Mineralization yielding about 0.50 per cent copper may be observed in the dump of the Deep Shaft distributed through Stage 2 quartz monzonite. Apparently similar mineralization is observable on the dump of the Wilson shaft.

These occurrences may indicate that Stage 2 alteration in the vicinity of more intense mineralization could possibly contain marginal copper values. On the other hand, the nuclei of ore deposition are associated with the more altered rocks. Also, the study of old workings suggests strongly that Stages 1 and 2, in both the alaskite and dacite, are not the host rock for any extensive ore accumulation.

Alaskite Stage 2, the most extensive rock type in the El Tiro area occupies much of the western portion of the mapped area. Along the east it is bordered, in some places, by a narrow rim of Stage 3 alaskite and also contains a number of small, scattered patches of the same high-intensity alteration. Numerous long, tenuous monzonite dikes cut the alaskite with some attendant alteration and mineralization. Whether these indicate a larger mass of monzonite below, and if so, how far below, can be determined only by exploration. In places, as in the June tunnel, occasional copper green stains indicate that mineralization may be associated with the more highly altered dikes. On the whole, however, the mineralization is weak at the levels where it may be observed.

An en echelon series of faults lies along the eastern margin of the alaskite, the so-called "alaskite front" of Kingsbury, Entwistle, and Schmitt (1941). Fissure and breccia zones along this front provided the locale for the copper mineralization from the Kurtz to the Hilda mines. Many of the workings are inaccessible, but some drifts and stopes are still open. West of the Bay State tunnel (100 level, Daisy mine) a stope about 20 feet across and several times as long is heavily mineralized with native copper. The host rock is alaskite Stage 3, and the native copper fills fractures both along the sheeting and at various angles. There is a similar stope about 100 feet away.

These two slopes lie beneath Mount Hope. The rock above is alaskite Stage 2. Possibly the rock not far below may be highly altered quartz monzonite. Presumably the native copper accumulated along the fractures in the alaskite not far above the monzonite contact. No evidence of native copper was observed on the surface or in the shallow tunnels.

The native copper has been relatively little oxidized. Some cuprite has formed at the expense of the native metal, and a small amount of malachite, but most of the mineral appears to be in its original state of deposition.

The alteration zone in the El Tiro area trends generally north. It is not so continuous as the east-west alteration zone of the Oxide deposit but has been more productive. The zone appears to extend almost uninterrupted from the Mammoth Canyon to the vicinity of the Kurtz mine. Localized alteration along fractures is observable (Pl. 3, fig. 3). Northeast of Jesuit Hill the zone continues through the dacite porphyry and quartz monzonite. Scattered spots of quartz monzonite Stage 4 alteration may be observed in a groundmass that is prevailingly dacite porphyry Stage 3. Indications of copper mineralization may be observed in the altered quartz monzonite.

Considerable areas of Stages 1 and 2 monzonite, and dacite porphyry are exposed along and east of the alaskite front. The area has attracted some mining as indicated by the Deep, Wilson, and Yankee shafts, as well as the Sampson tunnels. However, mineralization on the dumps and observation of underground workings indicate that this zone was by no means as attractive as the El Tiro zone in more highly altered rock.

Stewart (1912) has furnished a list of the minerals in the contact area and has outlined the geological relationships in the vicinity of the Mammoth mine by mapping the contact metamorphic zones.

Belt east of the Union Mine.—The Mammoth mine and the Union mine, together with the smaller Billy and Page mines, are reported to have had some 20 miles of underground workings. Judging from the association of the ores with garnet and limestone and a reconnaissance of the surface in the vicinity of the old workings, the mineralization belonged to what is now generally referred to as the contact-metamorphic type.

Although the copper mineralization accompanying the contact-metamorphic zones, particularly in the Mammoth mine, merits further investigation, no digression from the alteration

study has been attempted. A short time, however, has been devoted to the study of the surface along an alteration channel (Fig. 8) east of the Union mine.

This alteration channel was probably encountered in an inclined shaft next to the canyon bottom on the eastern side of the Union workings. Specimens on this dump near a caved inclined shaft with an eastward inclination contain disseminated sulphides including chalcocite and chalcopyrite rather than the usual contact-metamorphic waste material so common to the dumps of the Mammoth-Union area.

Quartz monzonite dikes which cut the limestone and contact-metamorphosed areas above the Union mine continue east through a broad area of dacite porphyry. These dikes are more highly altered on the hill east of Union Mine Canyon than within the area of the Union mine. The alteration also appears to coincide with a north-south alteration channel which varies from a few hundred to more than 1000 feet in width. The channel starts opposite but slightly south of the southernmost workings of the Union mine and extends northward for nearly a mile.

The chief rock type along the alteration channel is dacite porphyry Stage 3. Cutting the dacite porphyry at right angles to the axis of the alteration channel at intervals are quartz monzonite dikes which, within the alteration channel, are altered to Stages 3 or 4. There are also occasional patches of dacite porphyry Stage 4.

The alteration and the mineralization appear too weak to justify exploration. Mine dumps along the channel are quite highly mineralized, but the mineralization appears to be predominantly pyritic. A number of churn drill holes drilled in this area by the Imperial Copper Company apparently gave no evidence of more than sporadic showings of copper.

Alteration Minerals

General statement.—The alteration minerals are widely distributed but are seldom plentiful enough to provide a hand specimen consisting essentially of a single mineral; hence study of the minerals is almost entirely a problem of identifying mixtures. Small portions have been isolated for optical, x-ray, or thermal studies, but in many instances identification necessarily has been based upon microscopic methods.

The alteration minerals identified include the clay minerals kaolinite, illite (or hydromica), montmorillonite, and halloysite; the micas, sericite, and chlorite and the vein minerals jarosite, alunite, and barite, as well as opal, secondary quartz, and orthoclase or adularia.

Kaolinite.—Although kaolinite is widely distributed in the Silver Bell region where it occurs in small amounts as an alteration product of alaskite, dacite, or quartz monzonite, concentrations of comparatively pure kaolinite appear limited to zones of intense alteration in the alaskite. Here the mineral occurs in the late argillic phase associated with illite.

In a faulted zone northeast of Mt. Hope in the El Tiro area (Solano adit) a compact clay mass occurs as an advanced hydrothermal alteration product of alaskite. This material contains a small amount of illite but is largely kaolinite. The purified clay has indices n_γ = 1.568, n_α = 1.554 ± .002, and yields x-ray diffraction patterns which agree with kaolinite. The unpurified sample gives a characteristic thermal curve indicating kaolinite associated with minor illite.

At Butte, Montana, kaolinite forms a zone within the montmorillonite halo (Sales and Meyer, 1948), that is bleached white but contains specks of biotite. A second generation of this mineral in fine crystals appears to grow on the original biotite flakes.

Halloysite.—Halloysite is associated with alunite and jarosite in veinlets which penetrate altered alaskite along a ridge southeast of Wild Hog Butte. Specimens collected contain less than 1 per cent (estimated) of halloysite, so the presence of the mineral is not apt to be detected by thermal analysis or x-ray diffraction techniques, although it is visible in aggregates in thin section. The birefringence is extremely low with the index of refraction determined as 1.563 ± .002. This index is within the range for halloysite, although somewhat high which may be accounted for by intimate association with alunite. In general halloysite is a minor mineral in the alteration at Silver Bell.

Montmorillonite.—Montmorillonite with indices n_γ = 1.512 and n_α = 1.487 ± .002 has been found as one of the earliest clay minerals

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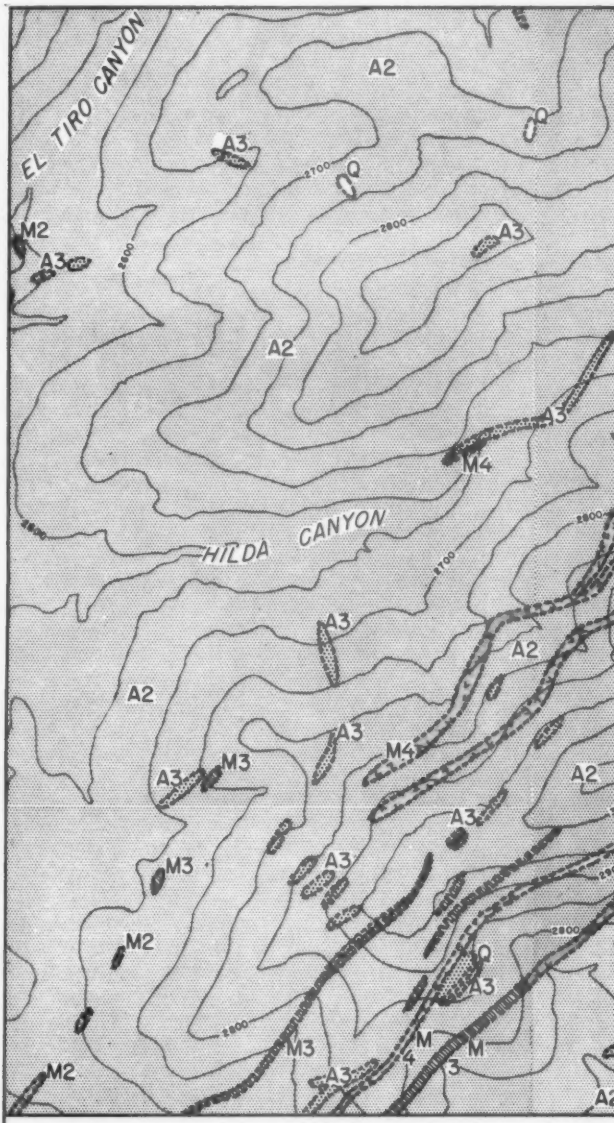
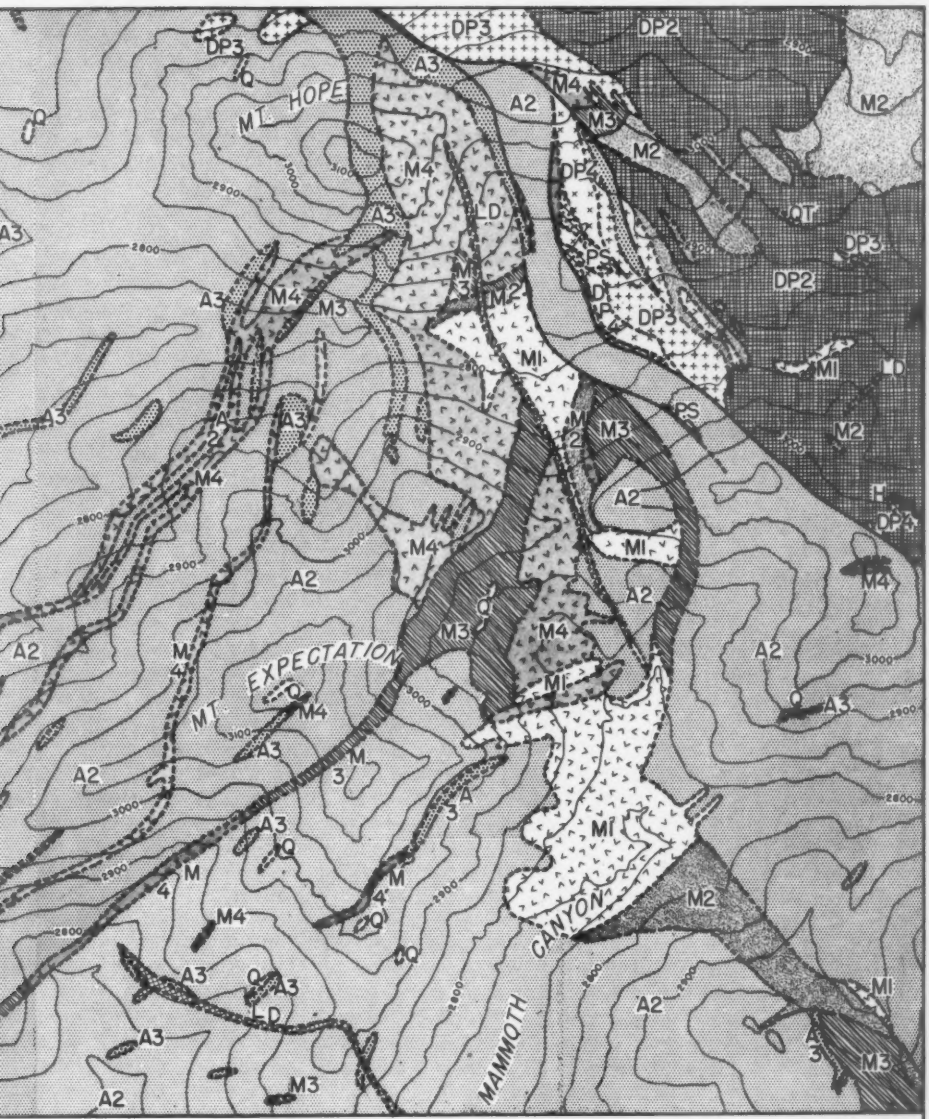


HYDROTHERMAL ALTERATION
IN A PORTION OF
THE EL TIRO DEPOSIT
SILVER BELL, ARIZONA

IGNEOUS ROCK
LATE DIKES
LD
QUARTZ VEINS-SILICIFIED
Q
QUARTZ MONZONITE P
M1 M2 M3
Alteration stages



US ROCKS
TE DIKES
LD
S-SILICIFIED ZONES
Q
ONZONITE PORPHYRY
2 M3 M4
ation stages

DACITE PORPHYRY
DP2 DP3 DP4
Alteration stages

ALASKITE
A2 A3
Alteration stages

SEDIMENTARY ROCKS
PALEOZOIC
PS
QT=QUARTZITE H=HORNFELS

Scale in feet
100 0 100 300 500 700

TAGES OF ALTERATION IN THE EL TIRO AREA

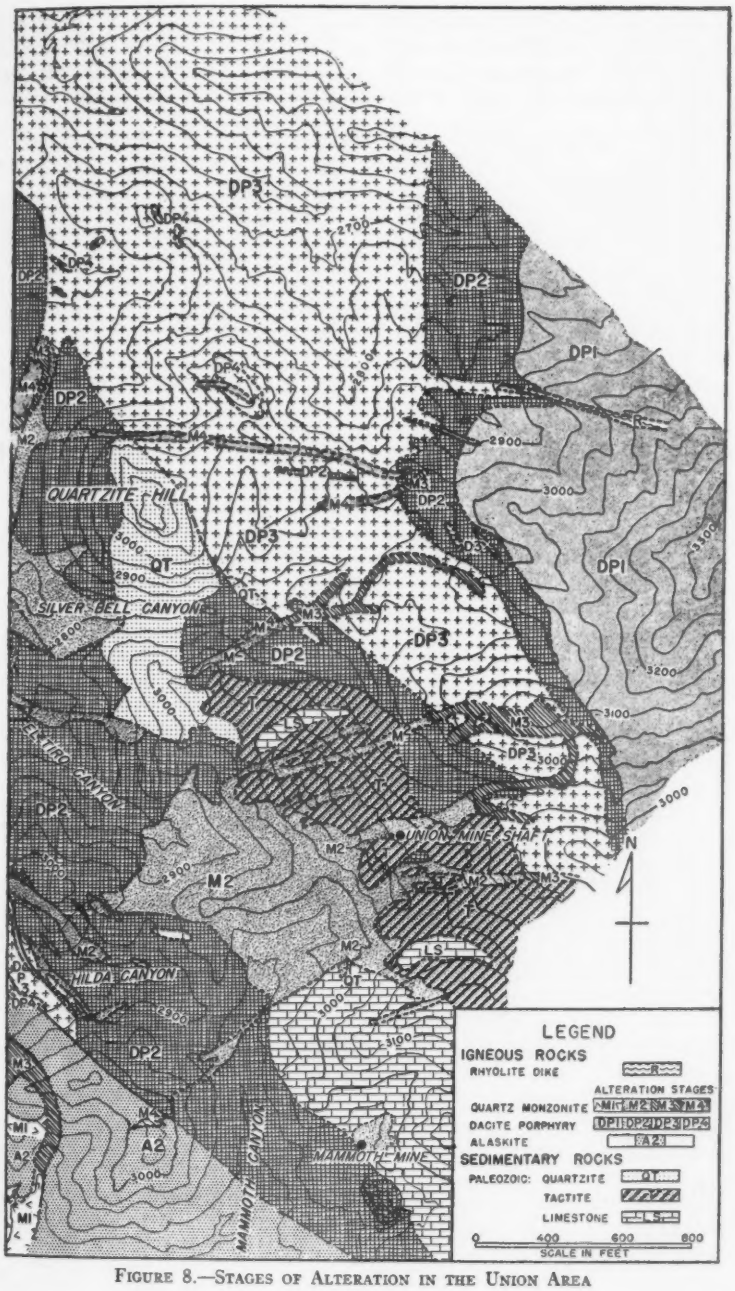


FIGURE 8.—STAGES OF ALTERATION IN THE UNION AREA

to form particularly in the alteration of dacite where it replaces small feldspar phenocrysts. It has been identified by its index, birefringence (observed .021, calculated .023), x-ray diffraction, and thermal analysis. It seems to be associated with illite rather than kaolinite, and no large concentrations have been found.

Although montmorillonite may form more readily in the originally more vitreous dacite, which is probably also more favorable as it contains more calcic feldspars, in general montmorillonite appears to follow the chlorite and represents the introductory phase of the argillic stage.

At Butte, Montana, in a careful study of wall-rock alteration Sales and Meyer (1948, p. 9) have identified a "montmorillonite type" clay mineral as the earliest argillic mineral to form. The mineral invades calcic plagioclase in particular either along cleavages or within more calcic zones. The thermal curve corresponds to curves previously recorded by Grim and Rowland (1942) for montmorillonite, the beta index is 1.53, and the x-ray diffraction patterns agree. The subzone carrying montmorillonite is the outer phase of the halo of alteration bordering veins on either side.

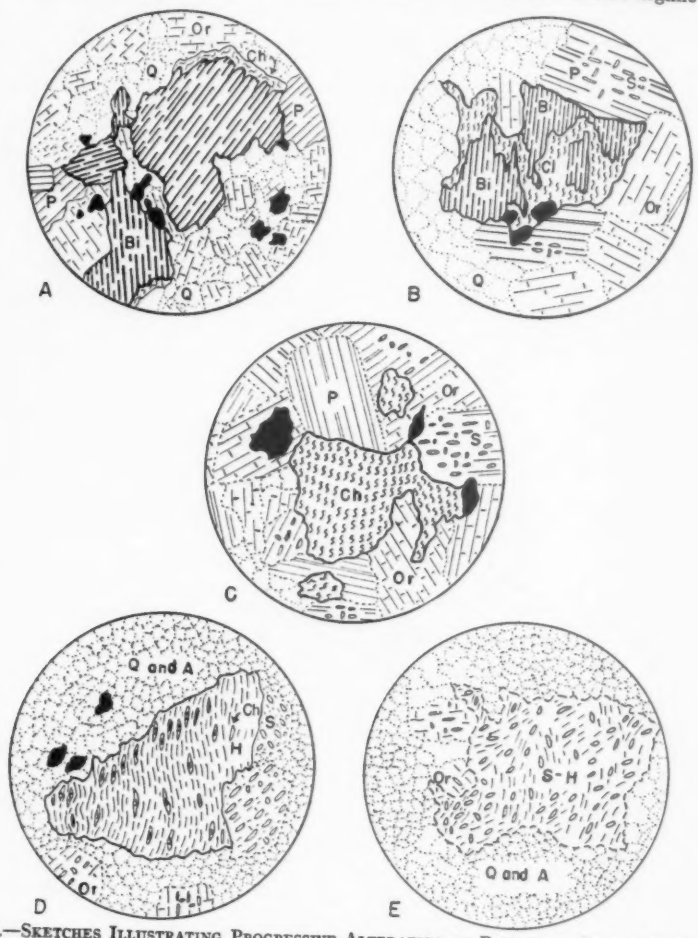


FIGURE 9.—SKETCHES ILLUSTRATING PROGRESSIVE ALTERATION OF BIOTITE IN QUARTZ MONZONITE

Illite (*or Hydromica*).—This material is be-

lieved to be one of the clay minerals of the alteration sequence. It appears to be formed at the expense of previous alteration minerals rather than from minerals of the igneous rocks. The clay mineral may be an advanced alteration product formed by the breakdown of alaskite, dacite, dacite porphyry, or quartz monzonite. It is not usually found in more than traces when chlorite is present, and it is replaced by quartz and sericite. It is likely to be preceded by montmorillonite and followed by kaolinite. All three, however, may occur close together in the alteration sequence.

The maximum index of refraction for illite in altered quartz monzonite from the north side of Wild Hog Butte has been determined as: $n_\gamma = 1.564 \pm .002$. The lesser index is low due to impurities. Another altered sample, dacite porphyry Stage 4, from the upper road on Copper Hill yields the indices $n_\alpha = 1.534$ and $n_\gamma = 1.568 \pm .002$. The observed birefringence for this specimen in thin section is .03, and the calculated value .034. The x-ray diffraction pattern is confused by the presence of some fine quartz lines, but additional lines due to the clay minerals agree with the patterns of illite. Thermal curves also agree with the thermal curves for illite.

These specimens have been compared with illite from the Fithian and Morris localities in Illinois (Kerr, Kulp, and Hamilton, 1949), originally described by Grim, Bray, and Bradley (1937) and recently discussed elsewhere (Kerr, 1949).

Sericite.—Fine white mica is probably the most widespread and most abundant mineral connected with progressive alteration. The x-ray diffraction pattern is similar to the pattern for muscovite, and the differential thermal curve is characteristically free from exothermic or endothermic reactions of consequence.

Several generations of sericite are shown by its replacement of feldspars and other minerals, but the major epoch in the alteration sequence occurs at the end of the argillic stage. Sericitization is also closely associated with later silicification. The thermal curve of an alaskite specimen from the second butte northwest of Copper Butte which is an area of intense sericitization shows a small amount of talc.

A central alteration zone at Butte, Montana, lies within the montmorillonite and kaolinite haloes (Sales and Meyer, 1948, p. 11). The central zone is characterized by sericite, quartz, and pyrite, in decreasing order of abundance; the sericite forms pseudomorphs after the earlier biotite. The quartz involved in silicification replaces sericite and may be distinguished from the original grains of the quartz monzonite.

Chlorite.—Almost every sizable area in which relatively slightly altered quartz monzonite predominates yields some chlorite as an alteration of biotite. This is particulalry true in a large area west of Copper Butte where Stage 2 quartz monzonite is widely exposed.

Chlorite may occur as a replacement of biotite and as an alteration mineral particularly in dacite. Greenish-gray masses high in chlorite are common in dacite in the southern and southeastern part of the area. Chlorite is detected by thermal analysis as well as thin-section study where the biotite gradually becomes less birefringent, pale green, and then is pleochroic to colorless. When fully developed it shows anomalous interference colors. Chlorite appears particularly vulnerable to secondary solutions, as it is often replaced by quartz, jarosite, alunite, or calcite. If not attacked by these mineralizing solutions, the chlorite apparently alters in the later stages to illite and in the final stage to sericite.

Alunite and jarosite.—Alunite and jarosite, ordinarily closely associated, are common secondary vein minerals in the Silver Bell region. In the Oxide area alunite and jarosite reach their maximum development where nearly pure specimens of the material are available. The concentration of either jarosite or alunite in various specimens may vary, but both occur together. X-ray diffraction patterns of the rather pure material indicate that the mixture is essentially heterogeneous and not an isomorphous series.

When *certain* fragments are observed in an index medium between alunite and jarosite, part of the fragment shows an index greater than the liquid, whereas the index on the other side is lower than the liquid. Such grains may represent ordinary mixtures or perhaps an edge of a zonal band in which part of a grain is jarosite and the remainder alunite. In thin section the alunite and jarosite appear as distinct minerals. Both are thermally reactive, and the curves are characteristic. Veinlets of alunite and jarosite cut most of the rocks of the region regardless of their alteration stage. The veinlets

cut the primary feldspars and quartz, as well as secondary quartz and sericite, and readily replace chlorite so the minerals are considered genetically later than most of the alteration effects. Apparently the alunite and jarosite are contemporaneous.

Barite.—Barite occurs in altered quartz monzonite on Copper Girl Hill and reaches its maximum development in veins such as those along House Canyon where large hand specimens of well-crystallized barite may be obtained. Barite is considered rather late; in thin sections veinlets of barite cut across altered quartz monzonite. The mineral has a characteristic bladed structure. The thermal curve of barite lacks exothermic or endothermic effects within the range of temperatures applied. In general, barite is not considered widespread in the area or significant in the alteration processes.

Opal.—Small opal veinlets in the rocks of the Oxide area occur associated with jarosite; apparently the opal follows the fractures partly filled previously by the jarosite. The invasion of the two apparently was not simultaneous since in several places opal cuts across jarosite. In thin section the opal is brown, has high relief, and no cleavage. The index of refraction was determined as $1.457 \pm .002$.

Secondary quartz and adularia.—During the late stages of alteration, a considerable amount of orthoclase or adularia and intimately associated quartz replaces large areas of the primary rock. The effects are striking since a fresh compact matrix forms in contrast to the highly altered feldspar phenocrysts the matrix surrounds, corrodes, and, in some cases, completely replaces. Occasionally, the adularia and quartz occur as larger grains replacing only a small amount of the highly altered rock so that the two appear to be present as fresh phenocrysts in a highly altered matrix of sericite, alunite, and clay. Secondary quartz unassociated with adularia is common and forms a network of veinlets in many rocks. It is particularly well developed in the more highly altered areas.

Accessory minerals.—Zircon, apatite, sphene, and rutile have been recognized as associated minerals. Small euhedral grains with the short prismatic habit common for zircon seem to persist throughout the various alteration stages. Minute prismatic apatite crystals are abundant, and many show hexagonal cross sections. Although apatite resists destruction in the early stages of alteration, as alteration progresses the crystals become more and more corroded, and in areas of most intense alteration the apatite disappears.

Sphene occurs most commonly in well-formed crystals having acute rhombic cross sections. This material is readily altered to leucoxene which often occurs as a pseudomorph which is characteristically yellowish white in reflected light. Rutile is relatively uncommon; it occurs as acicular crystals closely associated with quartz or occasionally as small grains or prismatic crystals.

Résumé

The alteration at Silver Bell in many respects is similar to hydrothermal alteration described by Lovering (1941; 1949a; 1949b), by Sales and Meyer (1948; 1949), by Schwartz (1947; 1949) and, by Peterson, Gilbert, and Quick (1946), and observed by the writer and associates at Santa Rita, New Mexico.

The most elaborate investigation of recent years was carried on by Lovering and his coworkers (1949a) in the East Tintic district, Utah. They have attempted to interpret the significance of hydrothermal alteration in considerably deformed Paleozoic sediments and overlying Packard rhyolite.

In their discussion of the Tintic District, Lindgren and Loughlin (1919) recognized three periods of alteration. The alteration which followed the close of volcanic activity accompanied ore deposition. In the near-by East Tintic District, Lovering (1949a) divides the alteration following the close of volcanic activity into: (1) an early barren stage, (2) a middle barren stage, (3) a late barren stage, (4) an early productive stage, and (5) a productive stage.

Stage (1) involved dolomitization of limestone and chloritization of the rhyolite. It would appear to correspond at Silver Bell essentially to Stage 2 in the quartz monzonite, dacite, and dacite porphyry and the early part of Stage 2 in the alaskite.

The second or mid-barren stage at East Tintic involved widespread argillic alteration in the rhyolite and localized argillic alteration

in the carbonate rocks. The principal minerals formed (Lovering, 1949a, p. 17) are a potash clay, halloysite, kaolinite, dickite, alunite, rutile, quartz, early hydromica, and early sericite. This stage would appear to correspond to Stage 3 at Silver Bell for the quartz monzonite, dacite, and dacite porphyry, late Stage 2 and early stage 3 for the alaskite.

The late barren stage at East Tintic is characterized by abundant silicification, pyritization, barite, and calcite. At Silver Bell silicification with pyritization appears to overlap and to follow sericitization and fall toward the end of Stage 4 for the quartz monzonite, dacite, and dacite porphyry, or toward the end of Stage 3 for the alaskite.

The early productive stage at East Tintic carries some quartz, abundant sericite, some hydromica, and pyrite and a little ore. The late productive stage is the ore-forming stage.

At Silver Bell the metallic epoch responsible for the introduction of the copper sulphides is earlier, corresponding to late Stage 3 and early Stage 4. Thus the early phases of the alteration at East Tintic and Silver Bell appear to be essentially parallel while the later stages differ.

The conditions at Silver Bell differ in a number of respects from the relationships described by Lovering (1949a) for East Tintic. This may be attributed to several factors and does not necessarily imply that different principles apply. The basic differences in the problem at Tintic would require a different approach. The Packard rhyolite obscures much of the underlying Paleozoic formations in which the replacement ore bodies are found. Thus the pattern of alteration in the rhyolite and its connection with mineralization become of fundamental importance. The resultant effects originating in a source below become affected in their distribution both by conduits in the Paleozoic sediments and fracture systems in the rhyolite.

Schwartz (1949) has emphasized convergence in metamorphism or the tendency for intense alteration to develop rocks that are remarkably alike. At Silver Bell convergence is so marked that it is difficult at places to distinguish between Stage 4 alteration of dacite, dacite porphyry, quartz monzonite, and even alaskite. This is particularly apparent in the most highly altered portion of the Oxide area.

Alteration phenomena somewhat parallel to Silver Bell have been described by Peterson, Gilbert, and Quick (1946) at Castle Dome. Here the quartz monzonite has been subjected to a sequence of mineral changes which involve weak propylitic alteration in the margin of the mineralized area, later argillic alteration, and finally a quartz-sericite stage.

At Santa Rita the hydrothermal alteration of the granodiorite porphyry is similar to the changes which have taken place in the quartz monzonite at Silver Bell (Kerr *et al.*, 1950). The parallelism is perhaps most evident in the biotite phenocrysts. Early chloritic alteration is significant at both localities. This is followed by a dominant argillic phase at each place. Sericitization and silicification constitute the final stages at both, and in each instance the early rock textures are largely obliterated during the process. Convergence is apparent at Santa Rita where in the final alteration stages the diorite sill, intrusive dikes, and the final stages in the granodiorite porphyry may be almost indistinguishable.

On the basis of assays in the open pit at Santa Rita, the argillic stage is the most closely connected with the copper deposition. The same relationship exists at Silver Bell.

Veins in igneous rock masses may provide an exceptional opportunity to study wall-rock alteration. An early description of such phenomena was furnished by Lovering (1941) in his discussion of the tungsten-bearing veins of Boulder County, Colorado. More recently Sales and Meyer (1948) have described the wall-rock alteration at Butte, Montana.

In each case selective replacement of wall-rock minerals by clay minerals has been reported, and a sequence of minerals significant of certain zones has been identified. Lovering (1941, p. 236) described an outer subzone where the chief alteration minerals are allophane and montmorillonite, a transitional zone with abundant beidellite, and an inner subzone with much dickite.

Sales and Meyer (1948) point out that every ore-bearing fracture regardless of its size, attitude, or relative age is surrounded by successive zones of sericitized and argillized quartz monzonite. A corresponding symmetrical sequence of wall-rock alteration is also attributed to each vein halo. In discussing the alteration along

the veins at Butte, Sales and Meyer (1948, p. 4) state that:

"the tentative conclusion must be advanced that the same solutions which deposited sulphide vein material were also capable of inducing contemporaneously both the envelopes of sericitization and argillization in the quartz monzonite wall rocks adjoining the channels through which they traveled."

This is of interest in view of the indications both at Silver Bell and at Santa Rita that the argillic alteration parallels the major concentration of sulphide mineralization.

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SEDIMENTARY BASINS OF ARIZONA AND ADJOINING AREAS

BY EDWIN D. MCKEE

ABSTRACT

An analysis has been made of the thickness of sediments accumulated in various parts of Arizona and adjoining areas during each of the periods of geologic history. From these and other sedimentary data, conclusions are drawn concerning the times and places of crustal movement in the region.

For the Paleozoic and Mesozoic eras, data have been compiled, period by period, in the form of isopach maps. For the Cenozoic era, information on thickness of sediments in individual basins is shown on a base map.

Five basic structures appear in Arizona on all of the Paleozoic maps. These consist of two positive areas (in the northeast and southwest respectively), the margins of two geosynclines (in the northwest and southeast), and a sag in central Arizona between the positive areas and connecting the two geosynclines. Prongs or submarine ridges extending basinward from the positive areas, and areas of more rapid sinking between them, appear to have shifted position from period to period.

Four major changes in isopach pattern are as follows: (1) The Uncompahgre-San Luis highlands of Colorado were formed and north central New Mexico became an area of more active and extensive sedimentation starting with the Pennsylvanian. (2) Central and southern Arizona became a high area contributing sediments to the north during Triassic time. (3) Northern Arizona received no sediments in Lower Cretaceous time, but in southern Arizona an uplift was followed by development of a deep, northwestward-trending trough from Mexico. (4) Deposits of Upper Cretaceous age were accumulated across northern Arizona in and near a sea that expanded westward from the Rocky Mountain trough.

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Introduction

This study of the distribution and thickness of sedimentary deposits in Arizona and adjoining areas is an attempt to assemble existing knowledge regarding times and places of sedimentation, and the relative rates of deposition within this region. Because areas of maximum accumulation ultimately develop into belts of intense folding and faulting and because such structures are closely related to igneous activity and to ore deposition, the study of sedimentary features is of great importance to an understanding of all other phases of regional geology. With this relationship in mind, data are presented here in a form designed to be useful for comparison with structural data presented in a paper in preparation by E. D. Wilson (1949).

The method employed in much of this investigation is essentially the same as that used by Ver Wiebe (1932) in his studies of the distribution and thickness of Paleozoic systems of the United States. An isopach map on which contours delimit areas of sedimentary accumulation in terms of hundreds of feet has been prepared for each geologic period for which data are available. Precambrian strata could not be treated in this manner because of insufficient information, nor could Cenozoic formations, because of irregular accumulation, mostly in valley basins.

Thicknesses of strata included within a system are obtained both from measured outcrops and from well sample data. Many of the figures for Arizona are from unpublished sections including measurements by the writer and his associates; others are from the literature. All of the data for adjoining areas have been taken from the work of other geologists. A key giving the source of each record follows the discussion.

The accuracy of data used doubtless varies for different parts of each map, depending on (1) the care with which measurements were made, (2) the structural simplicity of the sections involved, and (3) the completeness of the units represented. In some sections, the locations of boundaries between rocks of adjoining systems are open to question and may be subject to different interpretations, though it is doubtful that the general contour pattern for any map is greatly affected by errors of this type. The degree of accuracy may be said to be in rough proportion to the number of measurements available for the area.

Each figure used on the map represents, so far as can be determined, the approximate original thickness of the deposits of a particular system for a definite locality. Recognized exceptions are indicated by a (+) after the number. The assumption that a deposit includes the entire original accumulation is, of course, subject to error, but is made on the basis of evidence from overlying and underlying formations or from distinctive members or zones that indicate stratigraphic position. A zero is used only where both younger and older formations are present, suggesting that no deposits of the age indicated were laid down in that area. The unrepresented strata could have been removed by erosion during the time of hiatus, but, in general, local evidence will indicate the likelihood of such exceptions.

Stratigraphic records for some areas are largely obscure or nearly lacking because of structural complexities, metamorphism, or partial erosion. For this reason, where measurements cannot be obtained, but strata of a particular age are known to be present, these strata are recorded with a (+) only. Where both marine and continental deposits are in a section, their combined thickness is given in the isopach figure, except where specifically indicated otherwise. This is because the objective of the map is to show the total amount of accumulated sediments (proportionate to the total sinking) of a stated interval of time. In contrast, most paleogeographic maps attempt to show the boundaries of the sea at some particular point in time.

The location of contours around established thickness figures varies in reliability. Trends established in areas of abundant and well established data are projected into areas less well known and are paralleled by contours that are further removed. Through interpolation and projection most of the contour patterns are established, but in areas for which there is little information available, more than one interpretation may be possible. Such situations are discussed in the text wherever feasible.

Acknowledgments

This paper, accompanied by twelve maps, represents a compilation of data from many

sources including both published and unpublished sections, well samples and well logs. A large number of individuals and organizations have contributed toward its completion.

In the review and analysis of literature Messrs. Andrew Shride, Don Bryant, Paul Hughes, and Tom Parker, Mrs. Waldemar Bejnar, and Miss Katherine Dennison gave valuable assistance. Thickness of strata based on interpretation of well samples were furnished by C. D. Johnson, G. A. Hemenway, J. J. Wright, Robert Roth, H. S. Cave, and D. M. Ferebee.

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For critical suggestions on the manuscript, thanks are due Charles Anderson, S. C. Creasey, and Medora Kreiger, of the U. S. Geological Survey; Eldred Wilson of the Arizona Bureau of Mines; and B. S. Butler, University of Arizona. Helpful suggestions concerning the maps, by Charles Read and L. C. Craig of the U. S. Geological Survey, are also acknowledged with thanks. Drafting of maps was by Mr. Leo Bushman.

Previous Studies on Distribution of Land and Sea

The general distribution of land masses and seaways in Arizona and adjoining areas during most periods of history has long been recognized. Broad features of Paleozoic and Mesozoic time have been indicated on the paleogeographic maps of the United States by Chamberlin and Salisbury (1906), Willis (1909), Schuchert (1910), and Schuchert and Dunbar (1944). Recently, details of the Arizona Paleozoic have been illustrated in a series of maps by Stoyanow (1942). All of these studies have recognized the early existence of a major geosyncline (Cordilleran) forming the western border of the area in discussion and have indicated a land mass (Siouxia) extending southwestward from the Canadian Shield during much of the time. Furthermore, they have recognized a complete change of pattern in late Mesozoic with encroachment of seas from the east (Rock Mountain geosyncline) and destruction of the Cordilleran trough to the west.

Isopach maps of the United States by Ver Wiebe (1932) and paleotectonic maps of central and western North America by Eardley (1949) have confirmed many of the general conclusions previously reached on the basis of paleontologic stratigraphy. Also these maps have focused attention on areas of rapid sinking, in contrast with positive elements of little or no sinking. Equally important contributions but of more local character are the maps and descriptions of Burbank (1933) and Bass (1944) for southwestern Colorado, Read (Read and Wood, 1947; Read *et al.*, 1949) for New Mexico, and Nolan (1943) for Nevada.

Even a cursory examination of the accompanying isopach maps indicates that the pattern of contours indicating areas of rapid sedimentation and others of little or no sedimentation remained constant for the duration of several geologic periods before showing any notable changes. Thus, the pattern of Arizona and adjoining areas during early Paleozoic time applies with but little modification to all periods from Cambrian through Mississippian. A second slightly modified pattern applies to the Pennsylvanian, Permian, and Lower Triassic, whereas the Upper Triassic and Jurassic have a third. Patterns of Lower and Upper Cretaceous time are both distinctive, with features of distribution different from those of all other periods, making five major changes in the land-sea relations of Paleozoic and Mesozoic time in this region.

In addition to the record of rapid down-warping in some areas and of positive tendencies in others, data compiled on sedimentation also clearly indicate times and places of regional uplift. Evidence consists of large accumulations of coarse detrital sediments. Examples are found in the Pennsylvanian of Colorado, in the Upper Triassic of northern Arizona, and in the lowermost Cretaceous of southern Arizona. Less clearly defined as regards age but even more impressive from the standpoint of sediments accumulated, is the record of Tertiary and later crustal movements represented in the vast alluvial deposits filling deep valleys throughout much of the region.

Precambrian Basins of Deposition

A recent paper by Anderson (in press) on the early Precambrian rocks of Arizona, describing several areas in which detailed studies have been made in order to illustrate the structural trends and the lithologic characteristics,

Table 1.—Late Precambrian Strata and Thicknesses

Name	Locality	Thickness (Feet)	Reference
Uncompahgre fm.	S.W. Colorado	8,000	Burbank (1933)
Unkar-Chuar groups	east. Grand Canyon	12,000	Walcott (1895)
Apache group	southern Arizona	1,400	Ransome (1903)

brings out two features relating to early Precambrian (Archean) basins of deposition: (1) The predominance of north-south and northeast-southwest trending structures throughout the region, and (2) the tremendous thickness (more than 37,800 feet) of early Precambrian rocks represented in central Arizona.

Rocks assigned to late Precambrian (Algonkian) time are represented in many parts of the region. Unlike the older rocks, these show little or no metamorphism; therefore measuring minimum thicknesses for various areas is greatly facilitated. Within each area, some understanding of the character and environment of deposition and the influence of the basin on subsequent structures can be had. It is not yet possible to correlate with any real assurance the deposits of one basin with those of another, hence it does not seem practical to show these deposits on an isopach map.

Representative thicknesses of late Precambrian deposits in the Southwest area are presented in Table 1.

The relationship, if any, between these late Precambrian basins and succeeding ones of Paleozoic and Mesozoic age is at present obscure, though in some areas late faulting has been recurrent along zones of weakness that developed at least as early as late Precambrian time. Notable examples are the Bright Angel and Butte faults that extend across Grand Canyon and show clear evidence of movement at two or more widely separated times. (Walcott, 1890; Noble, 1914.)

Lower Paleozoic Deposits

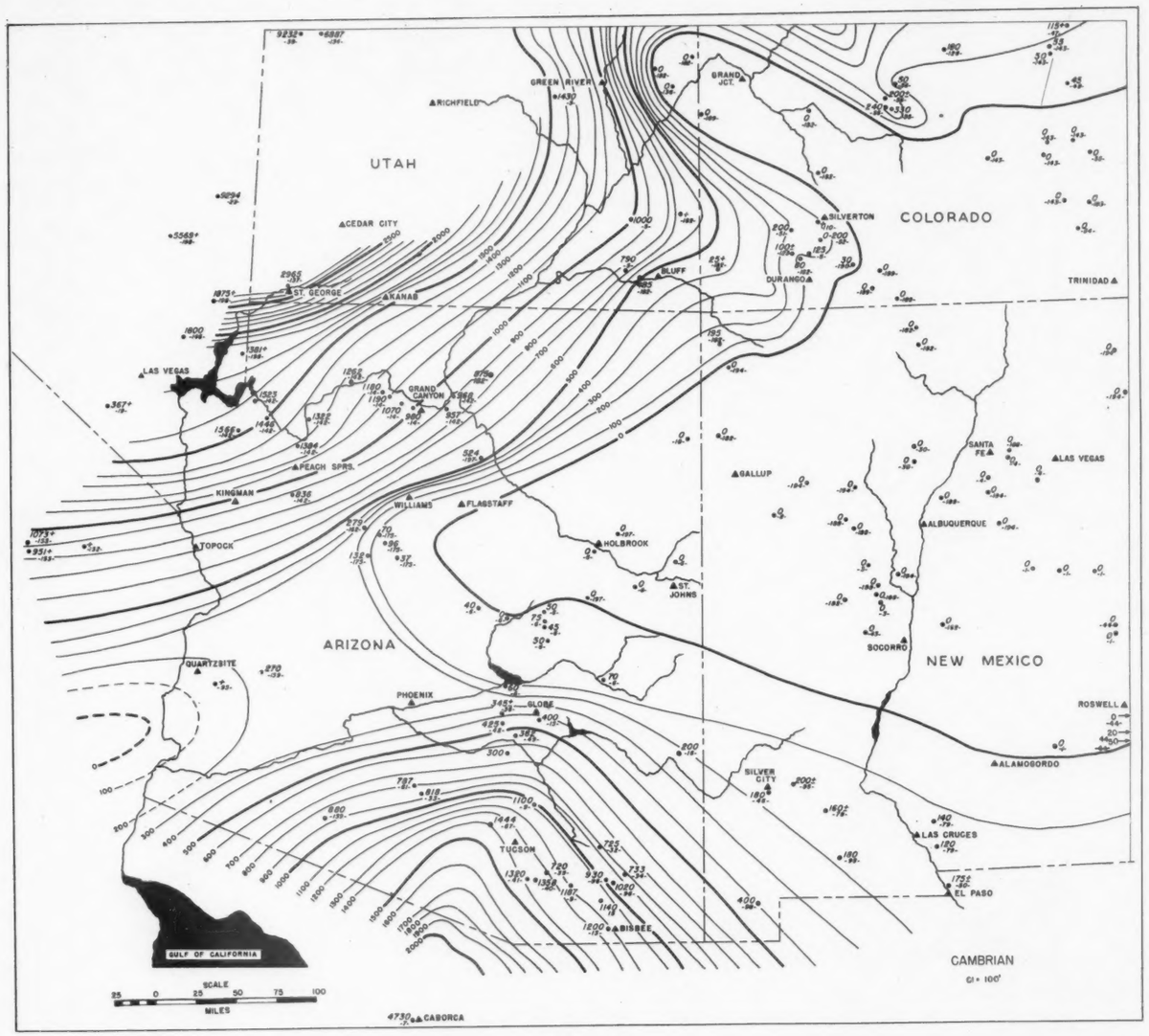
Pattern of Isopachs in Arizona

Isopach maps of Arizona during the Cambrian, Devonian, and Mississippian periods show many common features and only minor differences (Pl. 1, figs. 1, 2, 3). Total deposits of each of these periods in most parts of the state are measured in terms of a few hundred or less feet and do not exceed 2000 feet at any place. Ordovician and Silurian deposits are so limited in thickness and areal distribution in the Arizona region that isopach maps have not been made for them. Strata of the Ordovician period are confined to the southeastern and possibly the northwestern parts; those of the Silurian are not known within the state.

The most striking feature on all the lower Paleozoic isopach maps is the large area without deposits in the northeastern part of Arizona. It appears as a positive element, and has been referred to as the Defiance Positive Area. Lower Paleozoic sedimentary rocks increase in thickness away from it toward the south, west, and north. Eastward it is continuous with the large land mass that covered much of northern and central New Mexico from Cambrian through Mississippian time.[1]

The rate of thickening of total strata of each period is by no means uniform in various directions away from the Defiance Positive Area, nor are the areas of maximum thickening the same from one period to the next. Prongs or promontories formed by areas having thin deposits project westward or southward across Arizona between areas having greater thickness of sediments. In the Cambrian (Pl. 1, A), for example, the major prong extends westward from the region of Holbrook and lies considerably south of Grand Canyon. During the Devonian period (Pl. 2, B), a significant prong appears to have the approximate position occupied by Grand Canyon today, and a second one projects southward along the eastern border

[1] According to C. B. Read (personal communication) a thin blanket of Mississippian strata and possibly some Devonian, formerly included in Pennsylvanian system, cover large parts of north-central New Mexico.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RICHFIELD
CEDAR CITY
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DURANGO
TRINIDAD
LAS VEGAS
GRAND CANYON
PEACH SPRS.
KINGMAN
WILLIAMS
FLAGSTAFF
TOPOCK
HOLBROOK
ST. JOHNS
GALLUP
SANTA FE
ALBUQUERQUE
SOCORRO
QUARTZSITE
PHOENIX
GLOBE
SILVER CITY
ROSWELL
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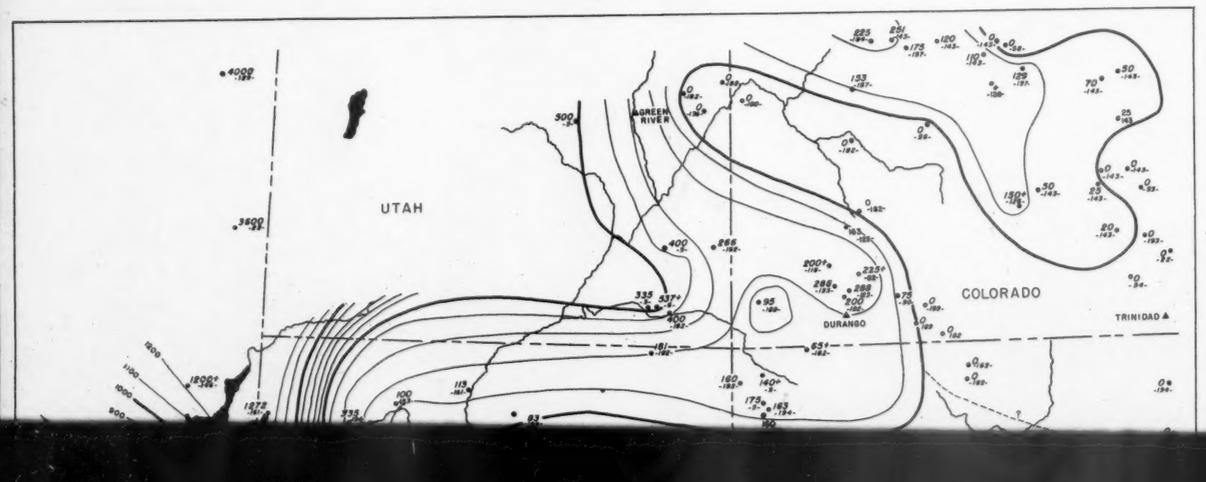
A CAMBRIAN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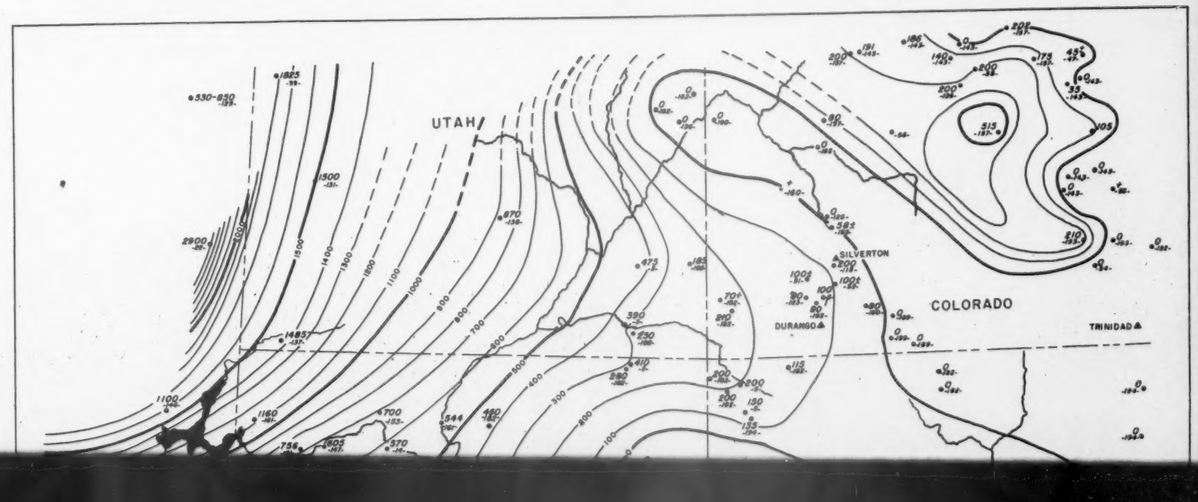
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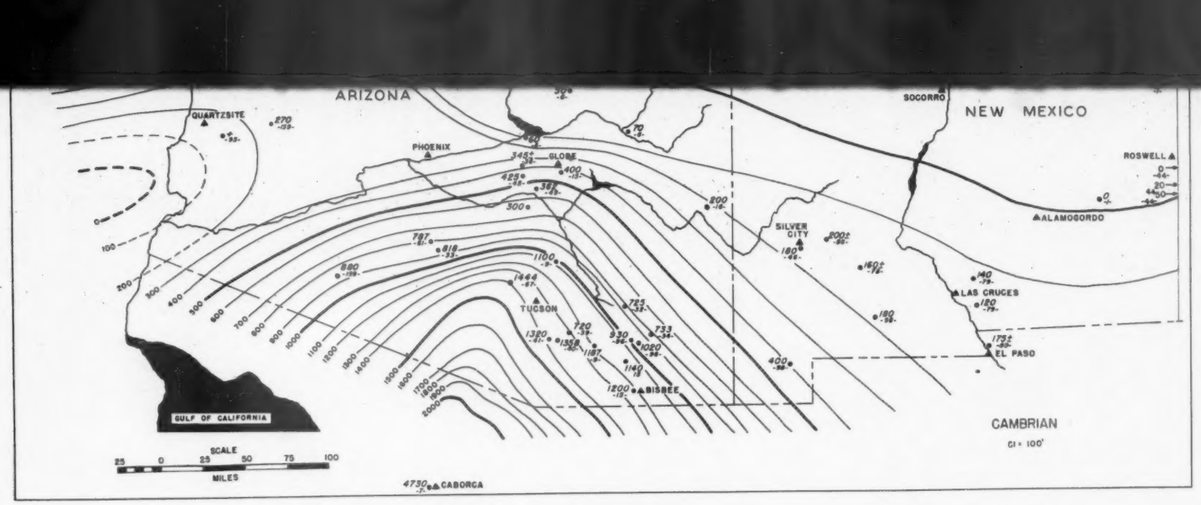

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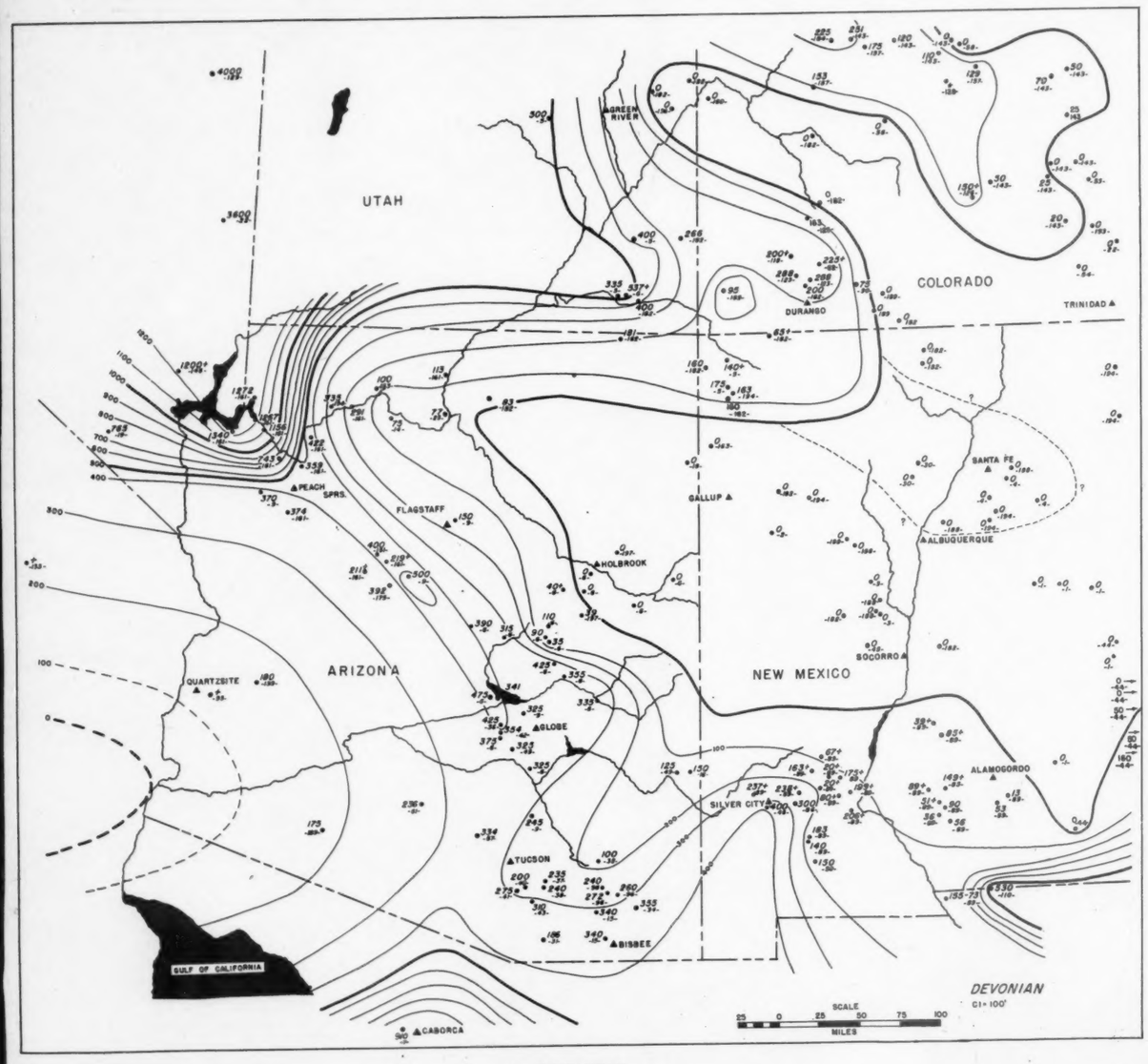
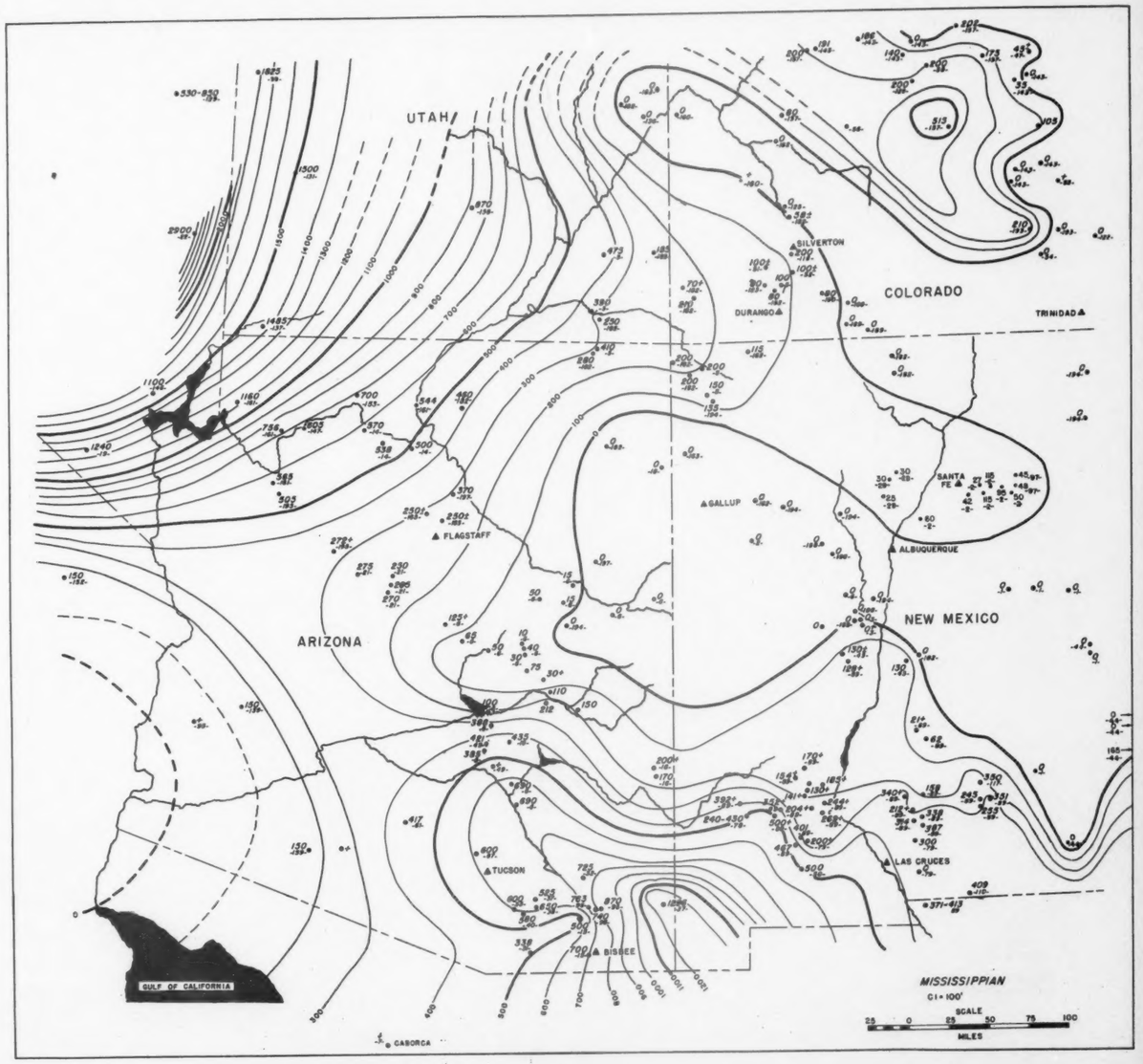


B DEVONIAN



C MISSISSIPPIAN

1. ISOPACH MAPS

of the state, with an area of comparatively thick deposits between. In the Mississippian (Pl. 1, fig. 3), a prong appears south of the Grand Canyon region, even further south than the prong of Cambrian age, whereas a belt of relatively thick deposits extends northward into the state from the southeast corner.

Prongs of early Paleozoic age formed by thin deposits and intervening basins with thick deposits probably represent belts of slow submergence and sedimentation between troughs of more rapid sinking and accumulation. An alternative explanation is that the prongs of Cambrian, Devonian, or Mississippian periods are areas that were upwarped and eroded at the conclusion of each period involved. Insofar as the writer knows, however, the character of sediments on the sides of the prongs gives no evidence to support this theory. Whether the prongs represent belts of uplift after deposition or the interprong areas result from more rapid sinking during deposition, the shifting in position of the prongs seems to oppose any concept of a single, well established submarine ridge of positive character across central Arizona.

A second positive element that affected early Paleozoic history in Arizona, as indicated by lack of deposits, is shown by zero contours in extreme southern California and adjacent parts of Mexico, not far from the southwestern corner of Arizona. Evidence for this land mass is not nearly as clear as for the Defiance Positive Area because, due to metamorphism and erosion, measurable sections and even recognizable outcrops of sedimentary strata are few in western Arizona and southeastern California. But by extending trends of contours and interpolation between a few well established points, existence of the positive element appears definite, though details of shape and boundaries are unknown. Especially significant in establishing the trends of contours in this area are recent unpublished data from the Harquahala and Growler Mountains obtained by Eldred D. Wilson (personal communication) and measurements from the Vekol Mountains by A. H. Carpenter (personal communication).

This positive element corresponds in general to Schuchert's Ensenada Land (1910, p. 470) and probably should retain that name. Eardley (1949) has shown this area as a region of "epeirogenic uplift subjected to erosion" on his paleotectonic maps of the Devonian and Mississippian periods (shown as a southeastward extension of the Manhattan geanticline during the latter). For pre-Devonian time, however, he indicates (1949, fig. 1) that this area was part of the rapidly sinking Cordilleran geosyncline and shows a 5000-foot contour crossing the southwestern corner of Arizona. Cambrian sections of rather considerable thickness in the Providence and Marble Mountains of the Mohave Desert, California, doubtless are the cause of his extending the basin southward. Those thicknesses definitely suggest a more southerly development of the geosyncline in Cambrian than in later periods; however, the thinness of the Cambrian section in the Harquahala Mountains (less than 300 feet), and the general southward thinning of the Cambrian in California incline the writer to believe that Ensenada Positive Area existed in Cambrian as well as in later periods of the Paleozoic.

Isopach maps indicate that during the Paleozoic era most of Arizona lying between the Defiance and Ensenada Positive Areas probably was a slowly sinking shelf, covered at various times with waters that connected the Cordilleran and Sonoran geosynclines. This shelf has been appropriately termed the "Arizona sag" by Eardley (1949, fig. 2). Contour patterns fail to indicate a Devonian landmass in this area such as the one Huddle and Dobrovolny (1946) describe as "a northeast-trending residue of pre-Cambrian mountains extending from southwestern Arizona through the Defiance Uplift to southwestern Colorado." Those authors present as evidence that this landmass existed in western and southwestern Arizona, an "increase in sandiness and an abrupt thinning" of Devonian sediments westward in central Arizona, but isopach data (Pl. 1, 5) do not substantiate this westward thinning. Likewise for the Mississippian, isopach contours (Pl. 1, C) suggest a trough trending from northwest to southeast across Arizona rather than a ridge connecting the Defiance and Ensenada Positive Areas. The westward-extending prong from the Defiance Positive Area might be interpreted as a submerged land mass, but this prong is not in the same position as those of Cambrian and Devonian time, so scarcely justifies classifi-

cation as a tectonic element of major significance.

It is important to realize that the isopach patterns of Arizona show total deposits of the basins for stated periods and do not illustrate the paleogeographic boundaries of a particular time or times. The Cambrian period, for instance, is represented in northern Arizona by deposits of Lower and Middle Cambrian epochs but in southern Arizona by strata of late Middle and Upper Cambrian age. Likewise, a marked thickening of the Mississippian strata in southeastern Arizona is due in part to the presence of Upper as well as Lower Mississippian in that area.

Degree of accuracy in various parts of these maps obviously varies to a considerable extent. Over much of northeastern Arizona, where Mesozoic strata conceal the record of Paleozoic formations, few data are available. Here the projection of the trends of contours, established through many readings in the Grand Canyon and through well sample data from southern Utah, has determined the positions of contours. A different interpretation of their position is plausible, however. They may have curved southward around a basin that extended from Utah into the Black Mesa area of Arizona. Drilling is necessary to determine this point.

Establishing contours in central and southwestern Arizona is still more difficult. Measurements of Paleozoic strata are few and largely incomplete here; consequently, location of contours is open to question in many places. Evidence that Cambrian, Devonian, and Mississippian seaways covered these areas seems reasonably good because (1) thicknesses and type of deposits at Vekol Mountains on the south, Providence Mountains, California, to the west, and Grand Canyon to the north all indicate lack of proximity to any land mass; (2) outcrops in the Chino Valley-Verde Valley area of east central Arizona suggest continuous deposition across this part of the state; and (3) outcrops in the Growler and Harquahala Mountains, as well as fanglomerates containing gravels of early Paleozoic age at Ajo and New Water Mountains indicate marine deposition in these parts of western Arizona. Only how great an accumulation of sediments occurred is open to question.

Former Southward Extension of Plateau Formations

Isopach maps of Paleozoic periods show a large area southwest of the Colorado Plateau in the region of Prescott, Arizona, where no data are available. Outcrops in this area consist largely of Precambrian granites and metamorphic rocks, with Tertiary or later volcanics locally forming caps. Except along the margins of the region, Paleozoic or Mesozoic strata are completely missing. That the sequence of Paleozoic formations represented in the Colorado Plateau formerly extended over this region, however, is suggested by several distinct types of evidence.

First, the thickness and lithologic character of Paleozoic strata immediately to the north and east of this area give no indication, except in the Cambrian deposits, of development on the margins of a basin. Strata of Devonian age thicken toward this area from both north and east. Those of Mississippian age are as free of detrital sediment as farther north, and, although thinner than at Grand Canyon, are several hundred feet thick. They increase in thickness toward this area from the east. Strata of probable Pennsylvanian age occur in the part of the plateau immediately to the north, and Permian beds are 2000 feet thick at their present southernmost margin. Lower Triassic deposits of the Moenkopi formation likewise indicate by their thickness (300 feet) at the southern edge of the plateau that at one time they extended considerably further in that direction. But there is no evidence that any later Mesozoic formations ever occurred in this area and the evidence of Late Triassic uplift, discussed in this paper, precludes the possibility.

The most convincing evidence that Paleozoic strata once covered to a considerable thickness the Precambrian complex of central Arizona is found in a gravel analysis of Pliocene (?) deposits scattered over the surface along the entire southern margin of the Colorado Plateau. Gravels in this category have been described from the Hualpai Plateau by Koons (1948) and from the rim of Sycamore Canyon by Price (1950). They have been observed in other localities as far east as Fort Apache. These gravel deposits in most places rest on an eroded surface of the Permian Kaibab formation, yet

they are composed of pebbles and cobbles derived from rocks ranging in age from Precambrian to Permian. As there is little possibility that such gravels could have been transported from the north, east, or west, they must have accumulated through the wasting of a large elevated region to the south. Thus, the present Precambrian crystalline rocks of the Prescott area may be considered the basement complex of a block or blocks raised higher than the more northerly blocks of the Plateau and subsequently stripped of all Paleozoic and Mesozoic formations.

South of the southern margin of this region for which no data are available, the sequence of formations is of considerable thickness and, so far as known, gives no suggestion of having developed as marginal sea deposits. In the Vekol Mountains, 60 miles southwest of Phoenix, Arizona, R. H. Carpenter (personal communication) has obtained the following measurements for Paleozoic strata: Cambrian, 629 feet; Devonian, 236 feet; Mississippian, 417 feet; Pennsylvanian, 415 feet. Farther west in the New Water and Harquahala Mountains (McKee, 1947), strata of considerable thickness of Cambrian, Devonian, Mississippian, Pennsylvanian (?), and Permian age are represented. It seems probable that, during most periods of the Paleozoic, deposits accumulated in many parts of central Arizona where today there is no record and, during some periods, they probably were continuous across the entire area.

Effect of Buried Hills on Formation Thickness

The earliest seas to advance across Arizona during the Paleozoic era covered a surface whose relief ranged from moderate to low. Hills of varying type dotted parts of the area and formed islands within the seas as the region was submerged. The size and abundance of these hills appear to have been controlled by: (1) the resistance of the Precambrian strata involved, and (2) the distance inland (eastward) from the Cordilleran geosyncline toward the Defiance Positive Area.

From the standpoint of an isopach map, the character of this Precambrian surface is significant. In areas where thicknesses of early Paleozoic strata are measured in tens or even in hundreds of feet, the presence of a moderately high, buried hill of Precambrian rock may cause a formation locally to have only half or a quarter of its normal thickness for that area. Where the total thickness of a formation is measured in terms of thousands of feet, the effect of a buried hill on any particular measurement is proportionately less. This explains many irregularities among the isopach figures for Cambrian and, in some places, for Devonian strata in Arizona.

Across the northern part of Arizona, where exposures in the walls of Grand Canyon are unparalleled for excellence, the late Precambrian erosion surface has been examined and discussed by many geologists (Hinds, 1935; Sharp, 1940; McKee, 1945). It is noteworthy that late Precambrian quartzites form the highest, steepest hills, some rising 1200–1500 feet above their bases. Monadnocks, both of granite and schist and of quartzite, are far more numerous in eastern than in western Grand Canyon. Where these buried hills project high into the Tapeats sandstone or into the overlying Bright Angel shale, the total thickness of Cambrian strata is greatly reduced, but elsewhere there is a constant trend of thinning from west to east.

In central and east-central Arizona, a surface of considerable irregularity, formed on Precambrian rocks, likewise affects locally the thickness of Cambrian strata. Furthermore, because the Cambrian deposits did not accumulate to as great a thickness here as farther north, some of the ancient hills project completely through them and into deposits of Devonian and Mississippian age. An excellent illustration of the latter is near Payson, Arizona, where large fragments of Precambrian quartzite are incorporated in Devonian dolomite (Wilson, 1939) and at Natural Bridge where both Cambrian and Devonian strata rest against a hill of quartzite (Ransome, 1916). Numerous other examples of hills buried by early Paleozoic strata are reported by Shride (personal communication) from the region farther east. At the American Cres Mine, south end of the Sierra Ancha mountains, the entire lower massive member of the Cambrian Troy quartzite is missing, presumably because of a high area on the depositional surface. In contrast, an additional thick member called the

Chedski sandstone is described (Burchard, 1931) from the base of the Troy near the junction of Oak and Canyon creeks, Ft. Apache Indian Reservation, and doubtless represents a low basin on the floor of deposition. Other illustrations are recorded.

In southern Arizona, local variations in the thickness of the basal Cambrian sandstone (Bolsa, Troy) suggest that there, too, an irregular, hilly surface extended across the region at the time Cambrian seas encroached upon it. Because of poorer exposures, evidence is not so apparent as farther north. Furthermore, the sea did not cover this area as early (Late Middle Cambrian rather than early Middle or Lower Cambrian as in northern Arizona). Nevertheless, the general landscape probably was not greatly different.

Pre-Devonian Disturbance in Central Arizona

Evidence of local structural disturbance between the close of Cambrian (Troy quartzite) deposition and the beginning of Upper Devonian (Martin limestone) sedimentation has been noted in a few localities in east central Arizona, but not elsewhere in the state. Shride records one locality near Highway 60 at Black River where dipping strata of Troy quartzite and underlying Precambrian formations have been bevelled so that the Martin limestone rests on successively younger beds across the area. Huddle and Dobrovolny (1946, p. 1205) state that at one place the Martin limestone "thins abruptly against a cliff of Cambrian Troy quartzite. . . ." These and similar observations suggest pre-Devonian deformation and erosion developed locally in the Arizona seaway. The forming of such unconformities may be confined to the area adjoining the Defiance Positive Area, for the marginal parts of any depositional basin are more favorable than other parts to local interruptions in sedimentation.

Character of the Positive Areas

The northeast and southwest parts of Arizona have been referred to as positive elements during Paleozoic time. Over these areas, deposits representing the various periods are either absent or relatively thin. This may be due to non-deposition or to erosion.

Available data on the Defiance Positive Area of northeastern Arizona furnish little conclusive evidence concerning its early history. Continential Permian deposits rest upon Precambrian quartzites in various places. Probably at no time during the Paleozoic was this positive area elevated as a highland or mountain area for pre-Mesozoic strata in surrounding areas contain few conglomerates, arkoses, or other coarse detrital sediments. On the other hand, the area may have remained at or somewhat above sea level during all of the Paleozoic era, as indicated by certain east-west trends in the sedimentary record of Grand Canyon.

Evidences that Paleozoic seas became shoal and ended along the margins of the Defiance Positive Area are as follows. First, Cambrian limestones are notably more argillaceous and much thinner in the eastern end of Grand Canyon toward the positive area, than to the west. Second, a marked increase eastward in the proportion of sandstone to limestone and a general thinning of the deposits is apparent in strata of Pennsylvanian (?) age in Grand Canyon. Third, in Permian formations, the complete absence of marine deposits on or near the Defiance Positive Area and the semi-circle formed by strata of near-shore facies west and south of the area are clear evidence of the former existence of a land mass in that region. The situation is less definite in rocks of Mississippian age, but the fact that progressive eastward thinning in Grand Canyon affects every member in the formation and is not attributable solely to a removal of upper beds may be significant.

The Ensenada Positive Area, southwest of Arizona, is much less well known than the Defiance Positive Area, though data from isopach studies suggest that it was of similar type. Definite statements concerning details of its character are not as yet justified because of a lack of field evidence. It may be speculated on the basis of meager information from Paleozoic strata in surrounding regions, however, that like the Defiance Positive Area this element had positive tendencies, but involved no mountain uplift.

The Colorado Strait

The stratigraphic record in Colorado indicates that from Cambrian through Early Mis-

sissippian time the pattern of seaways and land masses remained constant. Various geologists (Emmons, 1890; Hills, 1890; Burbank, compahgre) between two distinct seaways during Cambrian, Devonian, and Early Mississippian time, or (Baker, Dane, and Reeside, 1933,

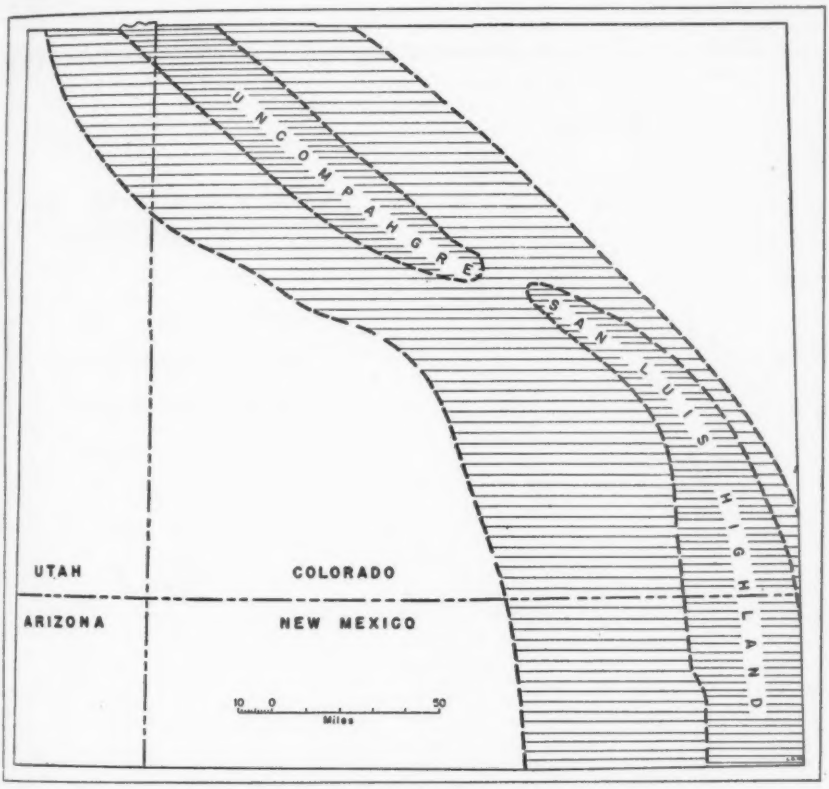


FIGURE 1.—POSITION OF UNCOMPAHGRE-SAN LUIS HIGHLAND

After N. W. Bass (1944), modified from Burbank (1933). Early stages indicated by area of closely spaced cross lines.

1933) have shown that a marine trough, bordered on the north by the landmass called Siouxia, extended from west to east across much of the state. The name "Colorado Strait" has been applied to this seaway (Ver Wiebe, 1932).

The position of the southern margin of the Colorado Strait is a controversial matter. Isopach maps show a long narrow belt or promontory (Fig. 1), outlined by a zero contour, that trends from northwest to southeast separating deposits of central from those of southwestern Colorado. This area of no deposits may have existed as a positive element (Unp. 975) the seas may have been continuous across this belt but their deposits in the area of Uncompahgre subsequently removed by erosion as a result of late Paleozoic uplifts.

Deposition of sediments in the Colorado seaways was small throughout early Paleozoic time. From Cambrian through Mississippian, a total of only about 1000 feet of strata was accumulated across central Colorado and an even smaller amount in the southwestern part of that state. Both in thickness and in character these deposits contrast strongly with those of late Paleozoic age in the same areas. They consist dominantly of carbonates and fine clastics,

whereas the later deposits include great thicknesses of conglomerate, arkose, and other coarse debris, clearly the result of crustal disturbances and mountain-building that occurred intermittently through the Pennsylvanian and Permian periods in the Uncompahgre area.

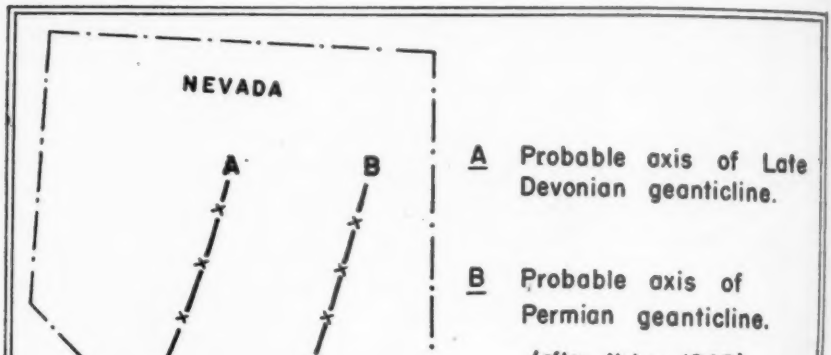


FIGURE 2.—AXES OF PALEOZOIC GEANTICLINES IN CORDILLERAN GEOSYNCLINE
After Nolan (1943).

Cordilleran Geosyncline and Geanticline of Early Paleozoic Time

The Cordilleran geosyncline, although outside the area investigated here, is so closely related to the problem that a summary of its early Paleozoic history seems desirable. In general this geosyncline extended from southwest to northeast—from southern California, through Nevada, into northwestern Utah. During parts of most periods, waters from this trough advanced over the more stable areas to the east, forming the seaways of northern and western Arizona and leaving a record of comparatively thin deposits. Whereas 30,000 feet or more of sediment, representing all of the periods and most of the epochs of Paleozoic time, were deposited in the geosyncline, only a few thousand feet of strata, and none of Ordovician or Silurian age, were laid down in Arizona during the corresponding time.

Nolan (1943) has shown that this trough was relatively simple from the beginning of the Cambrian period to the late stages of the Devonian; that during the close of the latter period a southwest trending geanticline or arch (Fig. 2) developed, splitting the geosyncline into two parts. Nolan also presents evidence

that, even when the geosyncline was a simple depression, areas of maximum subsidence, as indicated by thickness of deposits, shifted from time to time. Illustrations of this oscillation of the axis back and forth are (1) the Middle and Upper Cambrian in which the line of maximum thickness is some distance east of that for Lower Cambrian time and (2) the Lower Ordovician in which the axis is westward of that formed during Middle and Upper Ordovician time.

The geanticline that began to develop in late Devonian time was located approximately along the line that marked maximum sedimentation of the Cambrian and Ordovician periods. In dividing the old geosyncline into two new seaways, this arch forced the eastern margin farther east and shifted the zone of maximum sedimentation to western Utah and eastern Nevada. Other effects were the removal of early Devonian deposits along the axis of the arch and the accumulation of many shales and sandstones from erosion of the rising land. Although this geanticline lasted until Permian time, its greatest development was in the early Mississippian, for beds of this age are missing west of a line between Gold Hill, Utah, and Death Valley, California.

Of especial significance to the problem of Arizona seaways is the question of the southern extent of the geosyncline. Because of scanty outcrop data from southeastern California (the Mohave and Colorado desert areas), it is difficult to demonstrate whether or not the trough continued across this area and was a source of eastward floodings into Arizona. The trough may have terminated abruptly at the southern margin of the Great Basin or may have been deflected around it. Nolan (1943, p. 146) considers that trends in thickness and lithology as found in Nevada and adjacent areas, strongly indicate a former southward extension of the geosynclinal axis, and that the record has been partly obscured by metamorphism and partly destroyed by erosion. As already indicated, however, recent stratigraphic evidence from western Arizona indicates that a positive element persisted to the southwest of this state. Thus, any southward extension of the Cordilleran geosyncline would have been far to the west.

Pennsylvanian-Permian Deposits

Pattern of Isopachs in Arizona

The same five structural features that dominated topography and controlled erosion and deposition in Arizona during early Paleozoic time were important during the Pennsylvanian and Permian periods. Although the central Arizona belt of depression had been present during early Paleozoic periods, less than 600 feet of sediment accumulated in it during each period, whereas a greater amount developed during the Pennsylvanian, and over 2000 feet in the Permian. Regional uplift in central Arizona during or preceding Late Triassic time apparently destroyed the structure.

The Defiance Positive Area of northeastern Arizona, a prominent feature on all isopach maps of early Paleozoic time, continues conspicuous on maps of Pennsylvanian and Permian time. In the Pennsylvanian, this positive element became a small island because a sinking of the northern and central New Mexico area[2] caused encroachment of the seas from that direction. Later, during the Permian period, sediments actually covered the positive element though its outline continues to show on isopach maps in the form of contours surrounding a center of minimum deposition. The covering sediments of Permian age were entirely continental, doubtless resulting from extensive contemporaneous mountain-building in Colorado, whereas marine deposits nearly encircled but did not transgress the positive area.

Evidence that a positive area persisted to the southwest of Arizona (probably the Ensenada Land of Schuchert) is not nearly so clear as that for the Defiance Positive Area. Isopach data strongly suggest, nevertheless, that a significant land mass existed in that region. In western and southwestern Arizona, even though metamorphism and erosion have eradicated most of the sedimentary record, enough remains to indicate that deposition went on across the state during Pennsylvanian and Permian time, but was less in the southwestern than the central part of the state. Measurements by E.

[2] This sinking began in Mississippian or possibly earlier, as shown by thin deposits of middle Paleozoic age in parts of north central New Mexico (C. B. Read, personal communication).

D. Wilson (personal communication) during recent studies in the Harquahala and Growler Mountains seem to bear out this concept of a westward decrease in thickness. Measurements from the Providence Mountains (Thompson and Hazzard, 1946) and Marble Mountains (Hazzard, 1933) in southern California appear to indicate a southern end of the Cordilleran geosyncline in the Mohave desert area, lending further support to the idea of a positive element southwest of Arizona.

Crustal movements in certain areas outside of Arizona had a considerable effect on the physical history of the state. Especially noteworthy was the development of mountains in Colorado, described later. One range (Uncompahgre-San Luis ridge, Fig. 1) served to isolate a depositional basin in the southwestern corner of that state and to cause a very great thickness of detrital and evaporite sediments to be deposited there. Another development that took place in Pennsylvanian-Permian time was the general submergence, already referred to, of most of central and northern New Mexico. On the isopach map of Pennsylvanian time, furthermore, this area of submergence is marked by several areas of depression contours. The localities involved are ones of relatively little deposition, caused by monadnocks or hills of Precambrian rock that rose above the general peneplained surface on which accumulation of Pennsylvanian sediments began.

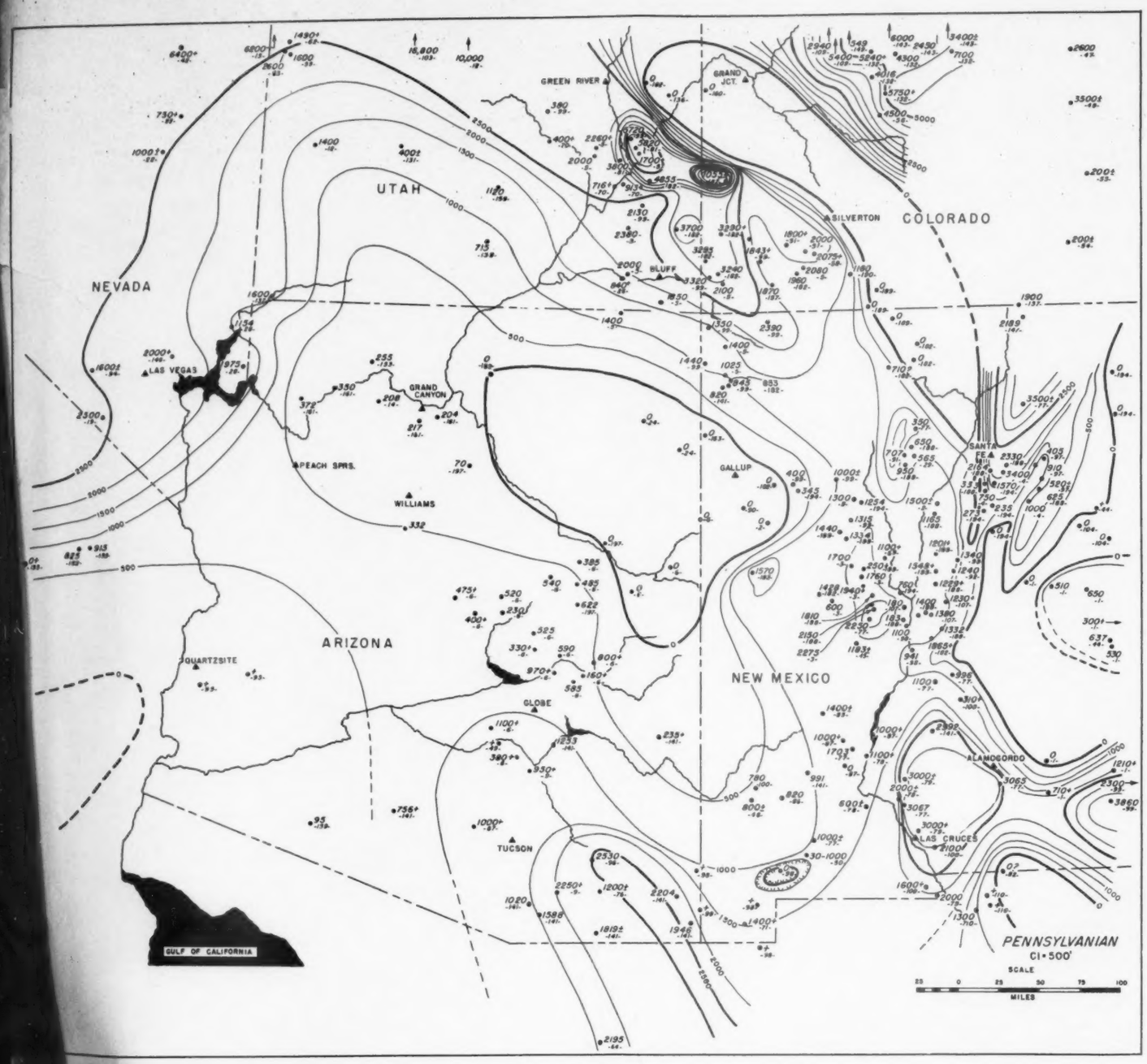
Basins of maximum sinking during the Pennsylvanian and Permian periods were in the Wasatch Mountain area, Utah (18,000 feet Pennsylvanian; 15,500 feet Permian) and in the west Texas area (12,000 feet, Permian), but neither of these was sufficiently close to have any direct effect on the Arizona problem. Reference to the Pennsylvanian isopach map (Pl. 2, A) shows basins of lesser magnitude but closer proximity, (1) north of Moab, Utah, southeastward onto Colorado (Paradox basin) 6000 to 5800 feet[3] (2) northeast of Santa Fe, New Mexico, 3500 feet; (3) between Alamogordo and Las Cruces, New Mexico, 3000 feet. Similar scattered basins of large sedimentary accumulation are found on the Permian map (Pl. 2, B), but not in the same localities. These include (1) south of Holbrook, Arizona, 3200 feet; (2) southeast of Socorro, New Mexico, 3300 feet; (3) southwest of Albuquerque, New Mexico, 2800 feet. The development of a positive element with no Permian deposits, in the area between Silver City and Las Cruces, New Mexico, is likewise noteworthy.

Certain errors in the preparation of the Pennsylvanian-Permian isopach maps appear to be unavoidable, especially with the present limited knowledge of the formations involved. One type of error results from the great thickness of the deposits which reduces the chances of finding complete sections. This problem, faced in all isopach work, but much less pronounced in the relatively thin deposits of the lower Paleozoic in Arizona, necessitates the use of maximum figures which in many places do not represent the original total thicknesses. A second type of error results from the difficulty of determining the boundary between Pennsylvanian and Permian strata. Everywhere in Arizona, insofar as the writer knows, there is gradational contact and a boundary is drawn within broad limits by datable fossils above and below an indeterminate zone. This method of establishing an approximate boundary has been used by Gilluly and Cooper in southern Arizona, by Huddle and Dobrovolny in east central Arizona, and by Longwell and Dunbar in southern Nevada. In much of the Grand Canyon area and farther south, however, no fossils are available so boundaries have had to be placed on the basis of lithologic types similar to those dated by faunas in adjacent areas. The method is subject to considerable errors but is the best that can be employed at present.

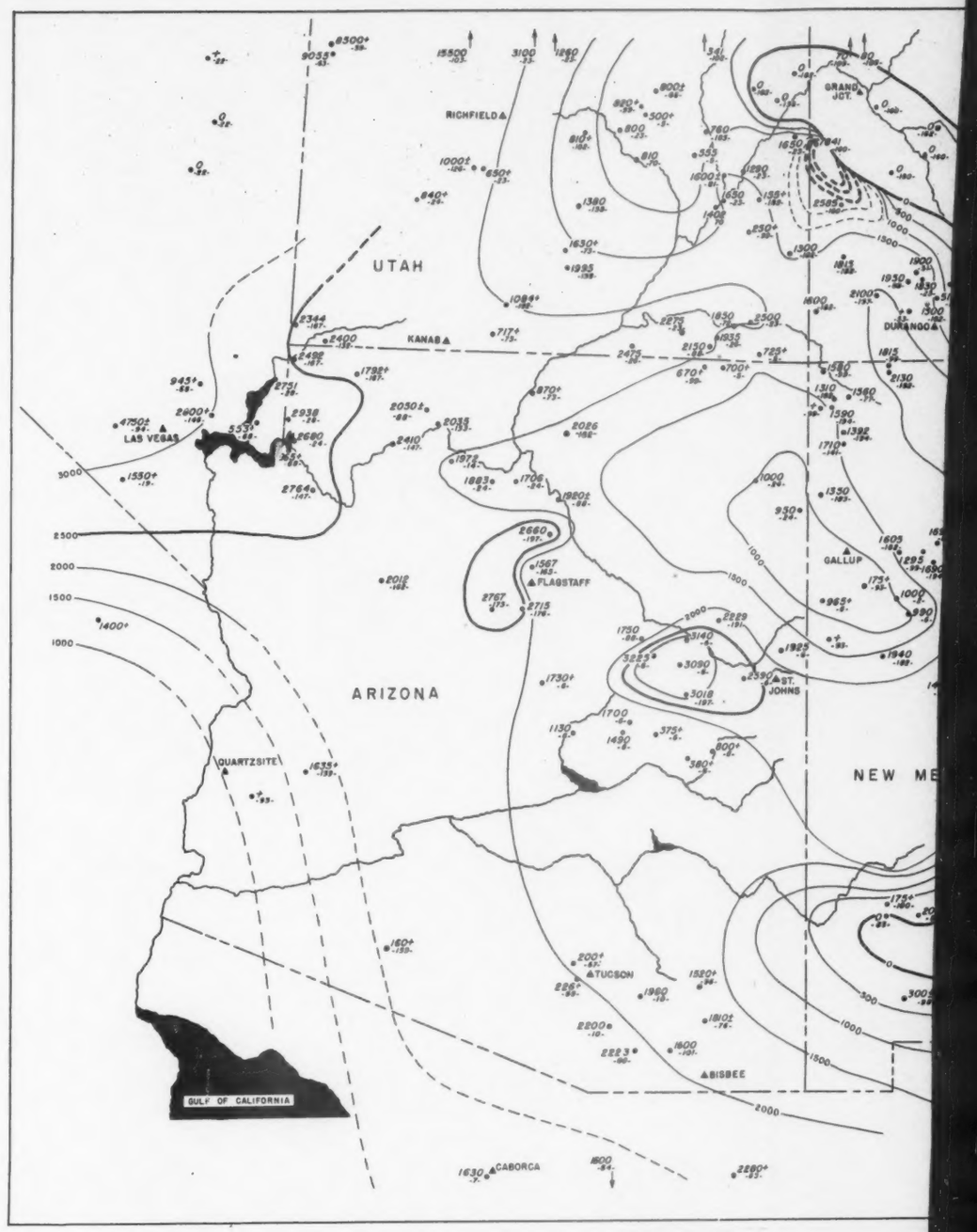
The Cordilleran Geosyncline during Pennsylvanian, Permian, and Triassic Time

The Cordilleran geosyncline, west of Arizona, appears to have been the site of progressively greater crustal instability and disturbance from late Paleozoic into early and middle Mesozoic time. A geanticline, formed during the late Devonian, separated the trough

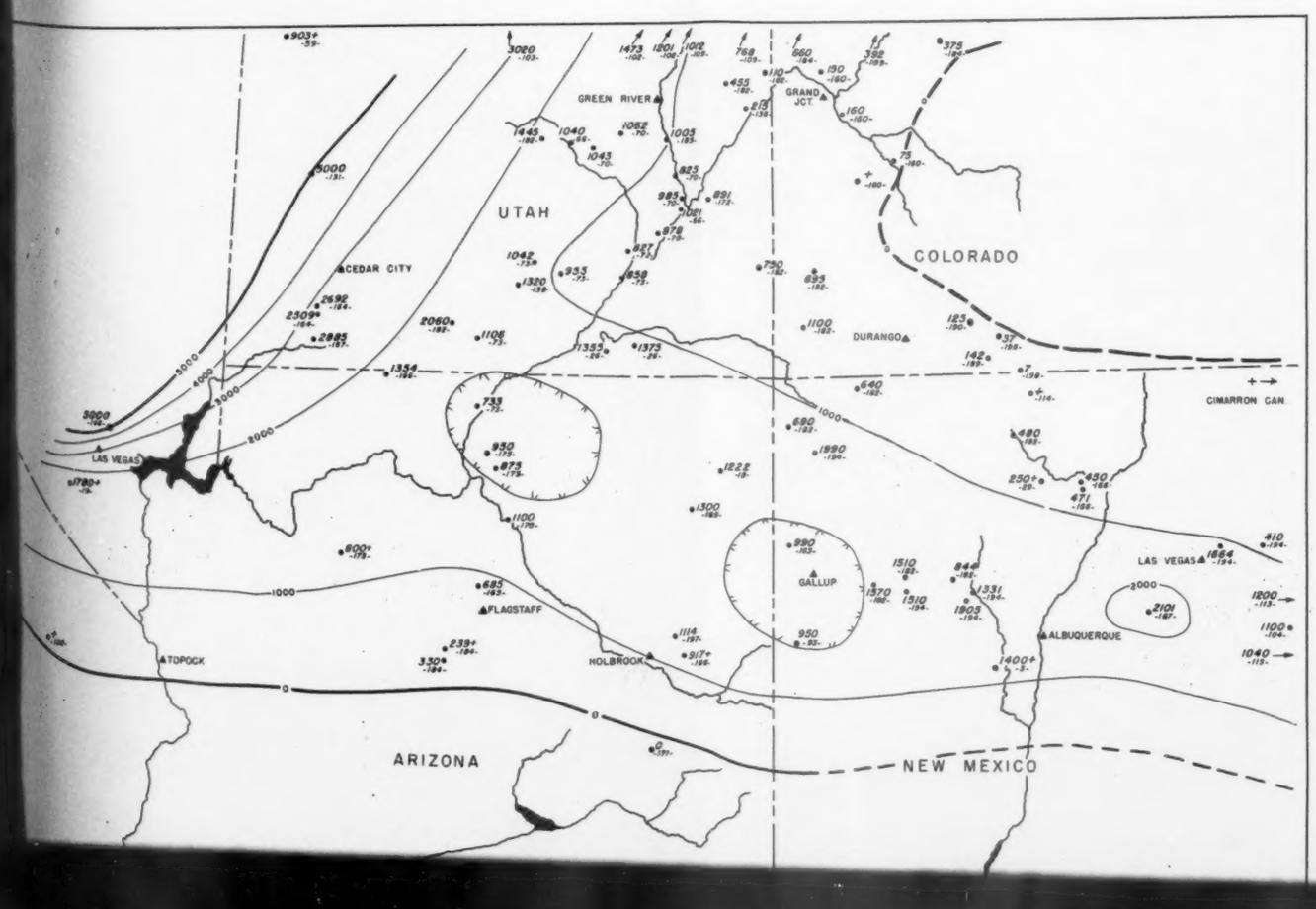
[3] These figures, based on well logs, probably greatly exceed original normal thicknesses of Pennsylvanian strata in the basin, as they result from lateral flowage of saliferous beds at anticlines. Strata of Pennsylvanian age probably exceeded 3500 feet in original thickness, however (Baker, Dane and Reeside, 1933).

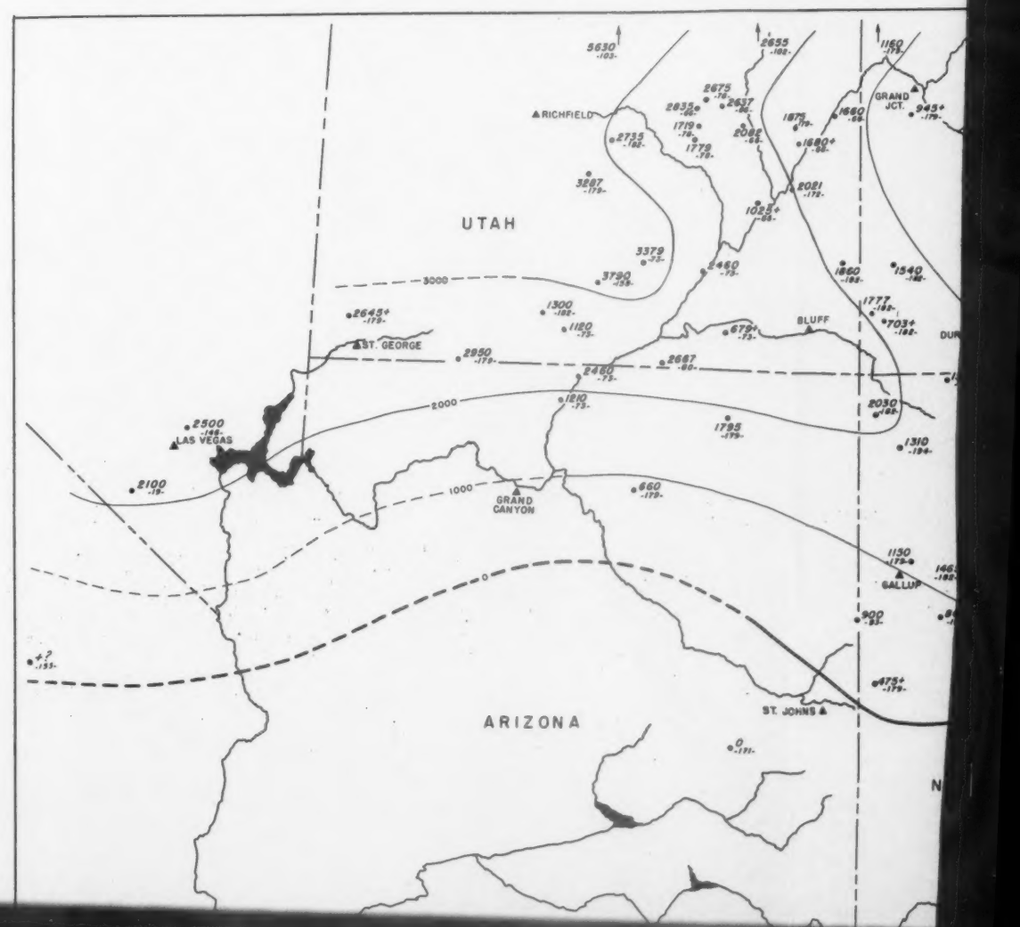


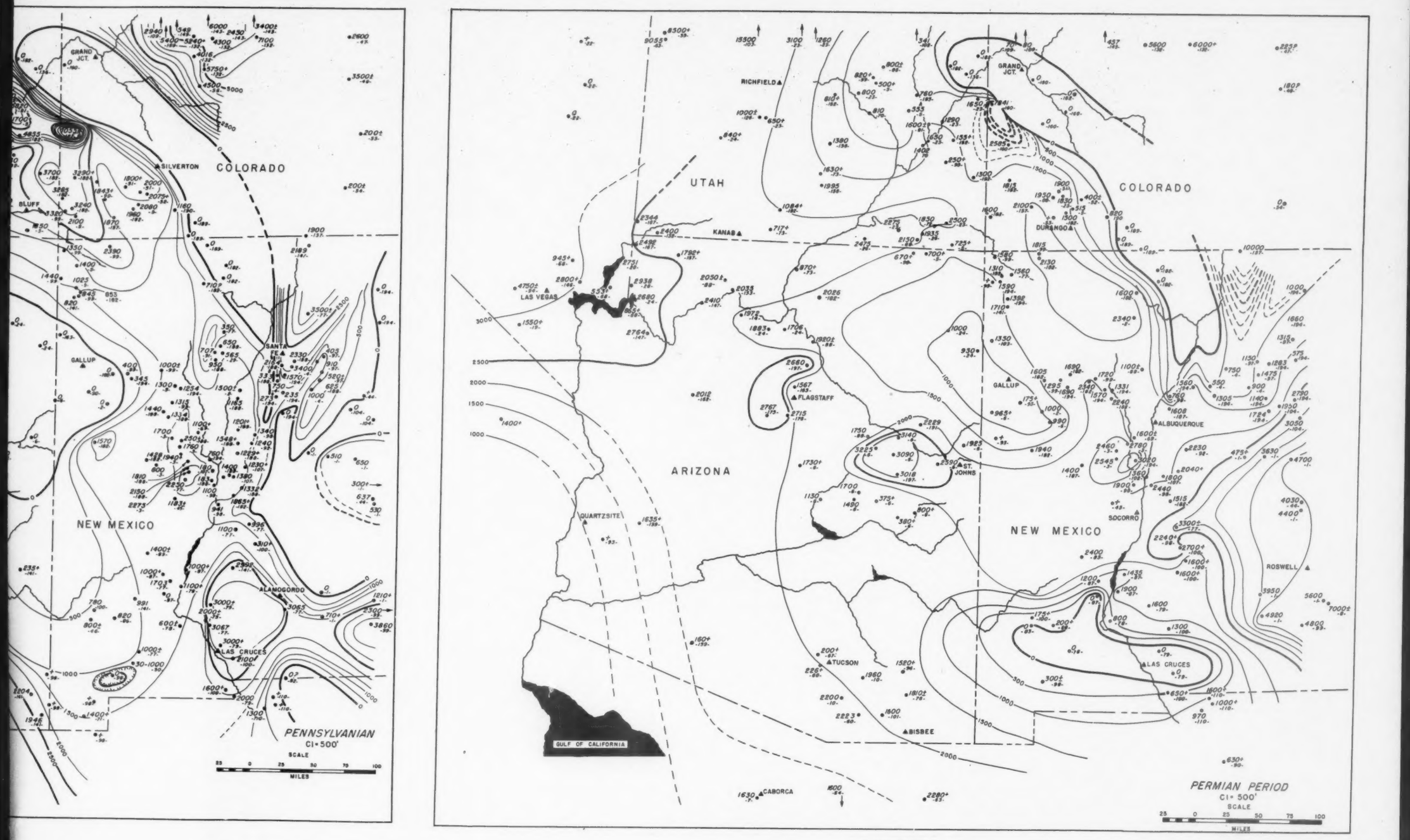
A PENNSYLVANIAN



B PERMIAN






GRAND JCT.
SILVERTON
COLORADO
BLUFF
GALLUP
SANTA FE
NEW MEXICO
ALAMOGORDO
LAS CRUCES
PENNSYLVANIAN
CI = 500'
SCALE
MILES
UTAH
RICHFIELD
KANAB
LAS VEGAS
FLAGSTAFF
ARIZONA
QUARTZSITE
TUCSON
BISBEE
GULF OF CALIFORNIA
CABORCA
DURANGO
ALBUQUERQUE
ST. JOHNS
SOCORRO
ROSWELL
PERMIAN PERIOD
CI = 500'
SCALE
MILES

B PERMIAN

GRAND JCT.
COLORADO
DURANGO
CIMARRON CAN
GALLUP
LAS VEGAS
ALBUQUERQUE
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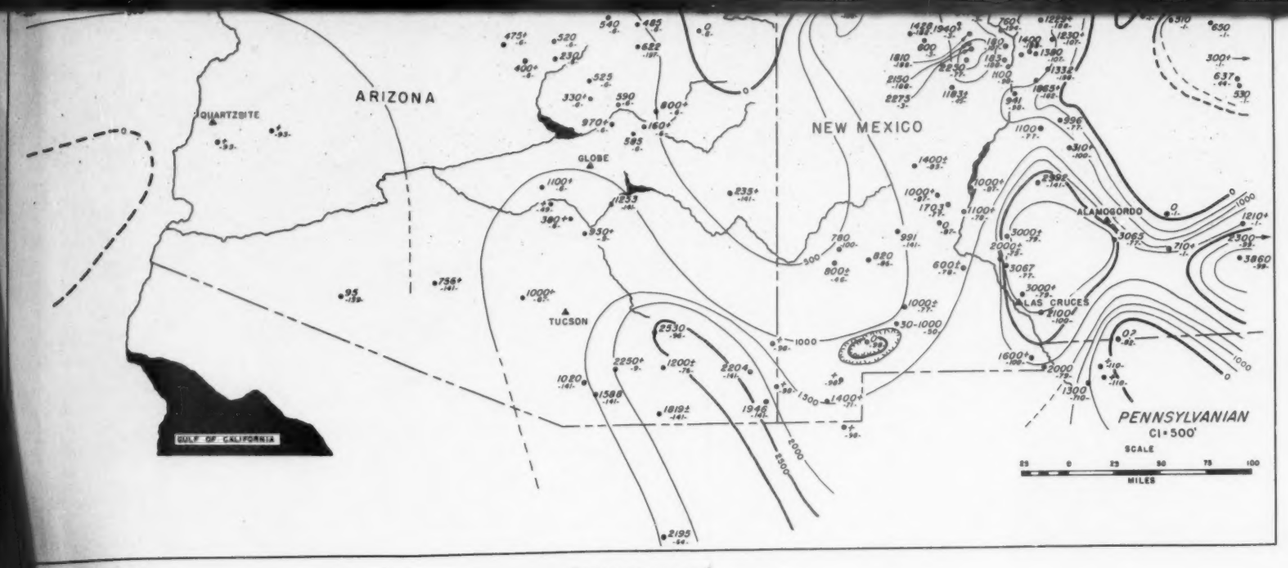
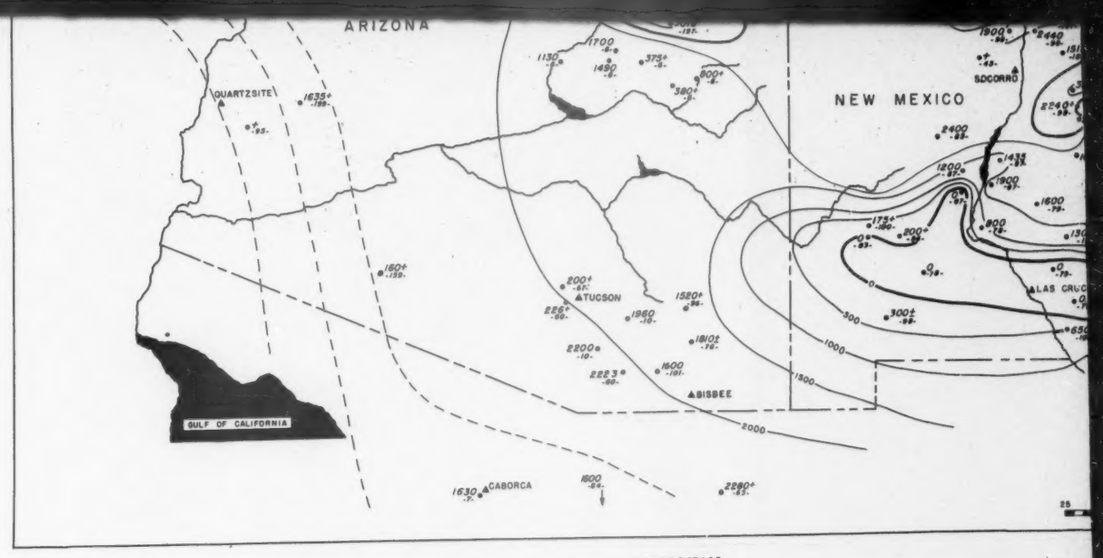


A PENNSYLVANIAN



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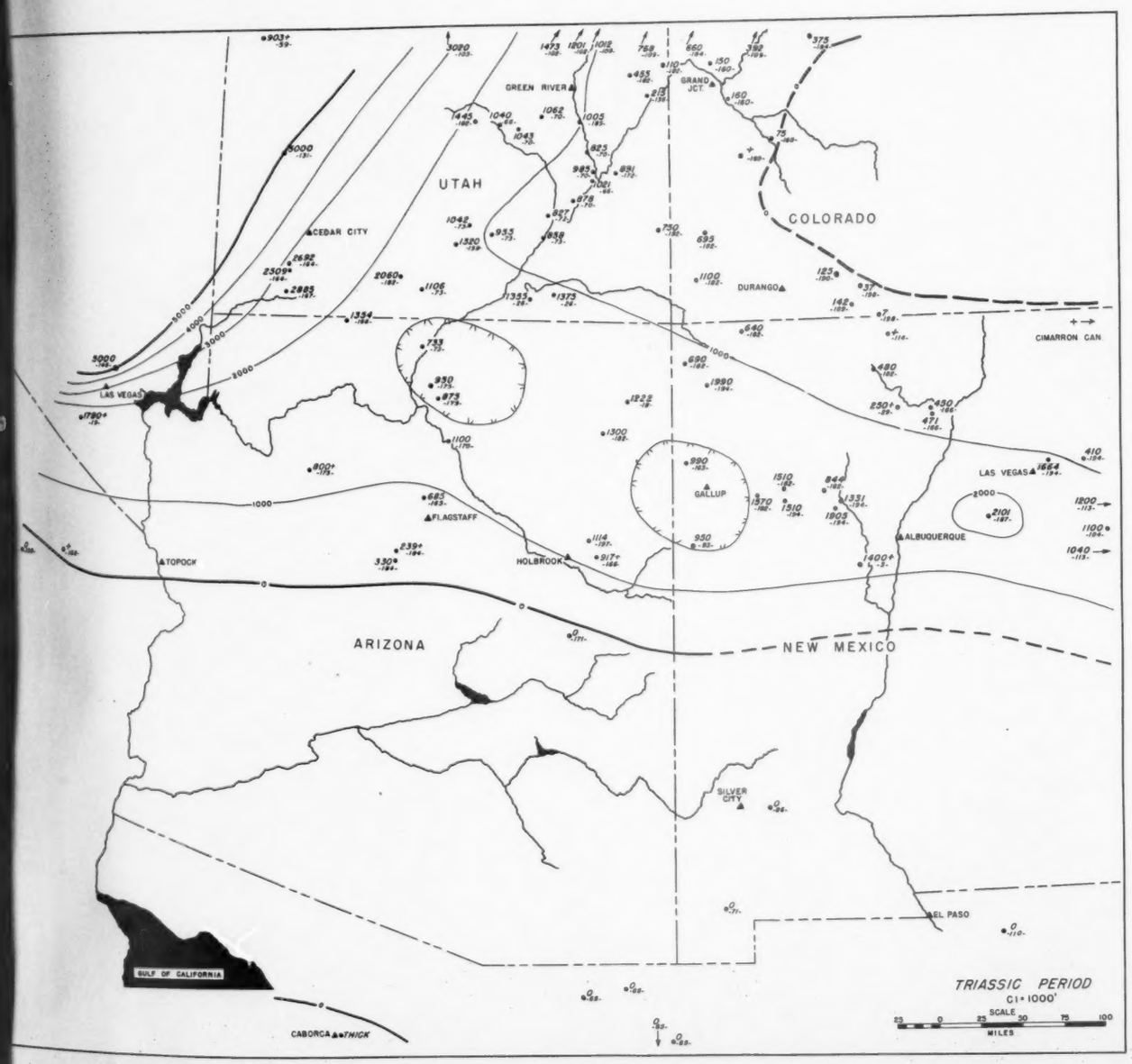
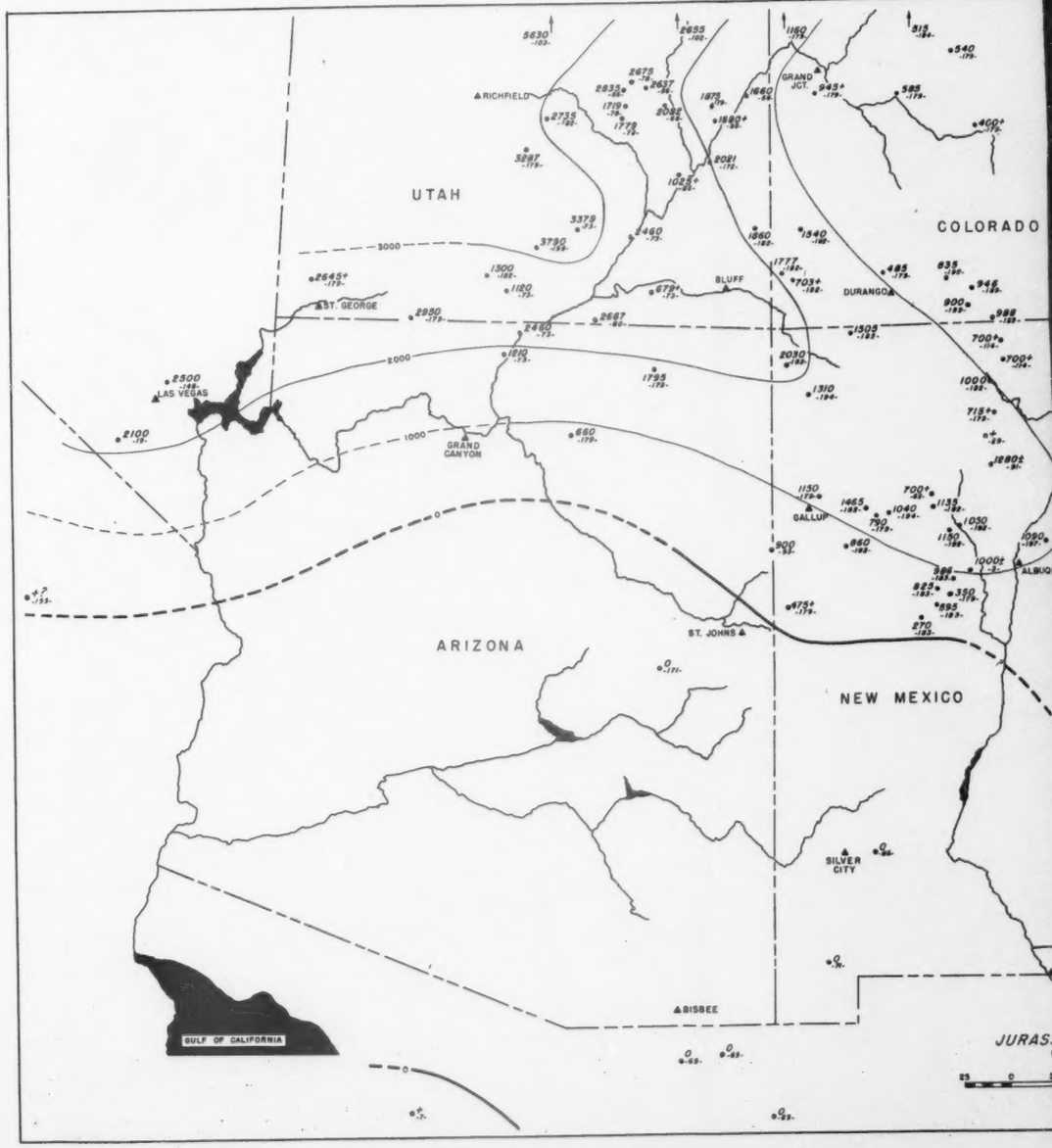
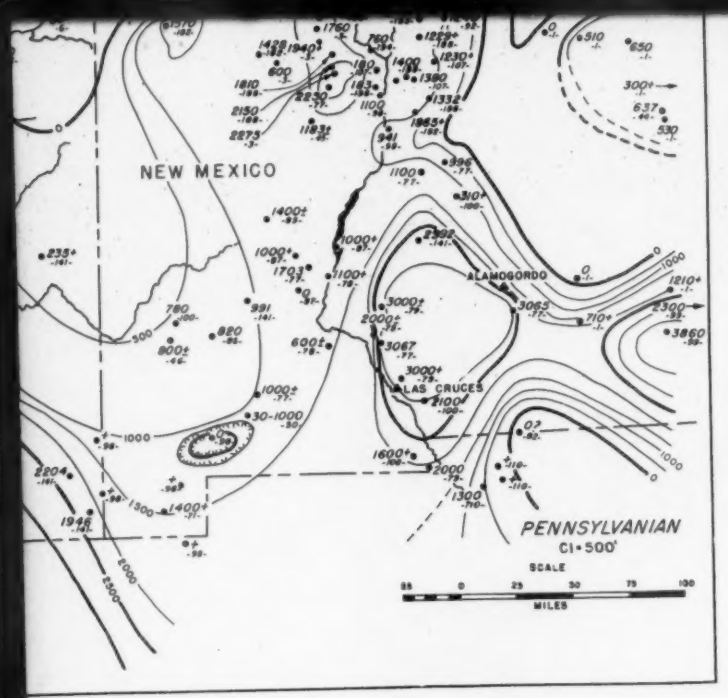


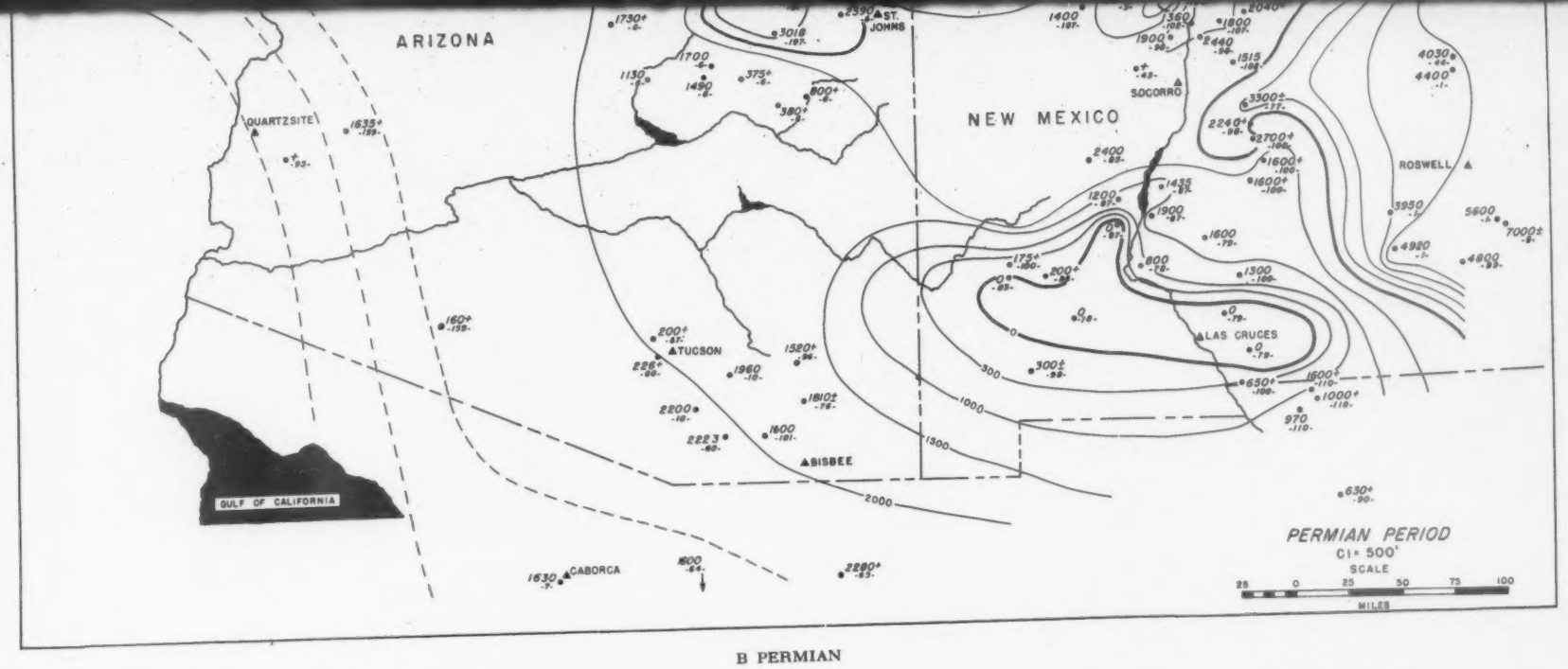
C TRIASSIC



D JURASSIC

2. ISOPACH MAPS





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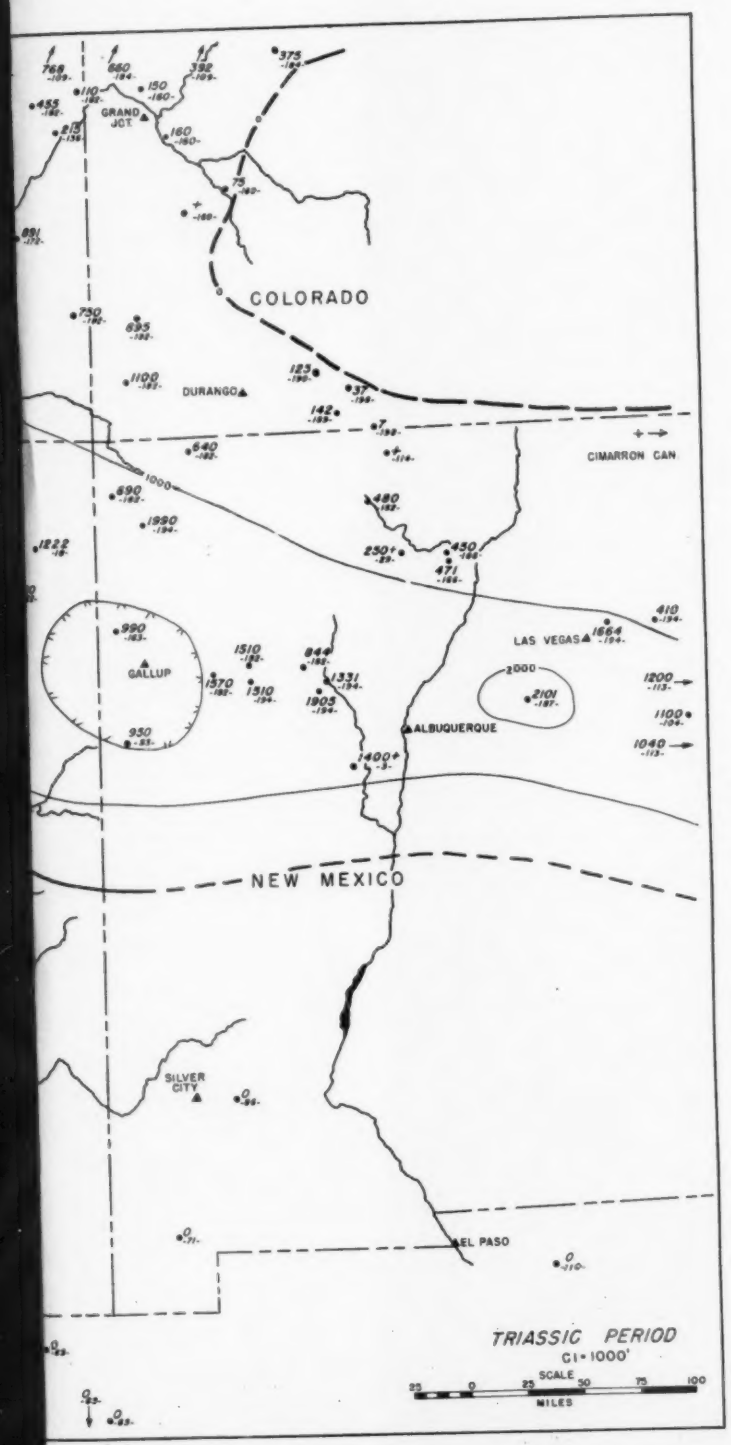
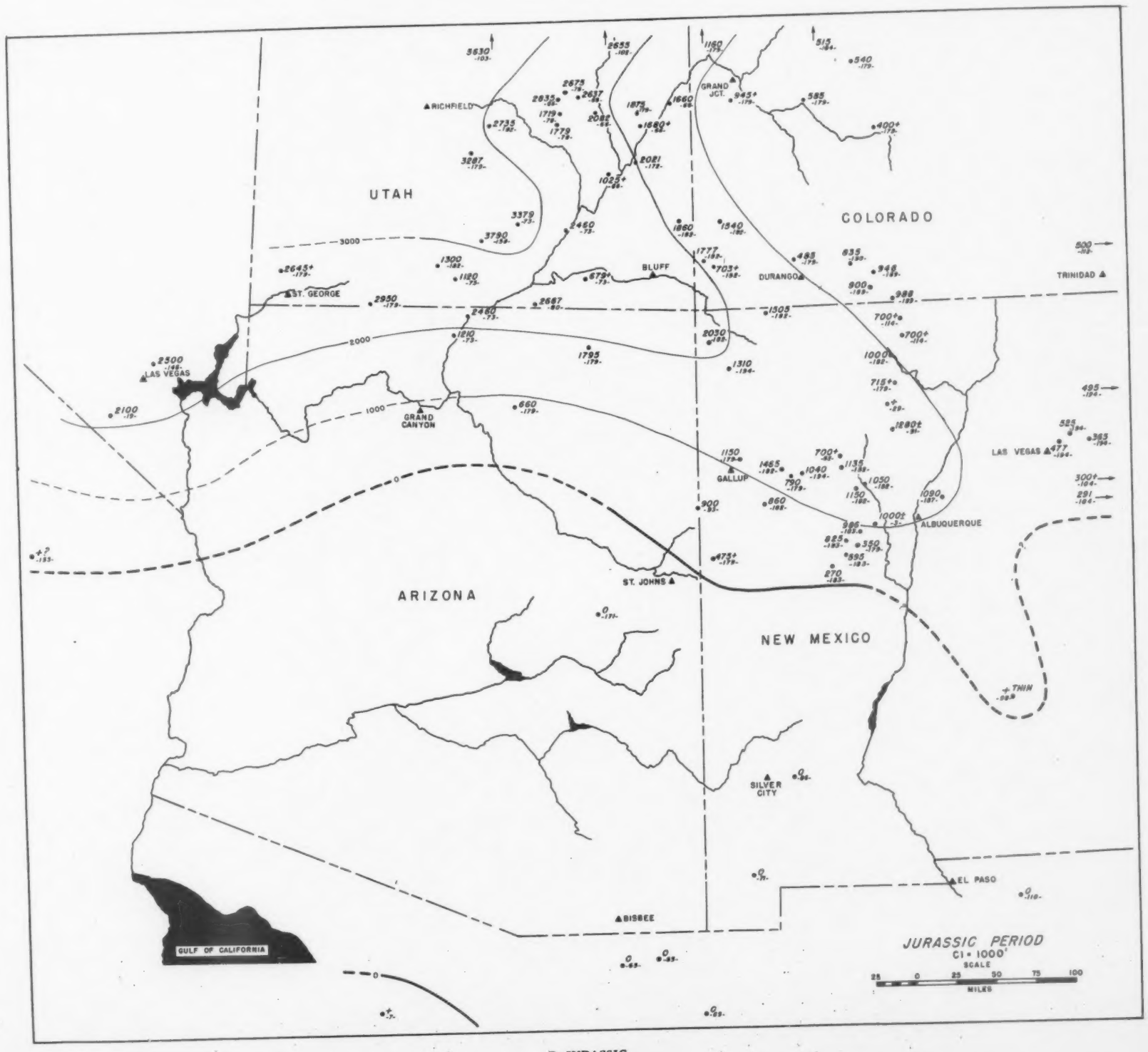




D JURASSIC

2. ISOPACH MAPS

into two parts until the close of the Pennsylvanian, and during that period the area east of the axis became very unstable. This is illustrated by Pennsylvanian sediments which (Nolan, 1943, p. 155) show (1) abrupt changes in thickness within short distances, (2) a high proportion of clastic materials, and (3) many local unconformities.

By Permian time, the first geanticline had been worn away (Nolan, 1943, p. 157) and thick deposits began to be laid down on its site. Almost simultaneously a new arch with its axis 100 miles farther east started to rise. The upwarping must have been gentle but progressively moved northward until by Late Triassic it extended entirely across Nevada. This second geanticline appears to have formed an effective barrier, because both faunas and lithology on the two sides were very different during much of this time. The eastern trough had its outer margins forced eastward into Arizona and Utah, and marine sediments merged into continental in that direction. The western trough was characterized by the accumulation of much volcanic material and, during Triassic, by very thick deposits.

Colorado Mountains of Late Paleozoic

Although no evidence of crustal disturbance in late Paleozoic time is known from Arizona, a considerable amount of uplift and mountain building took place to the northeast in Colorado. As described by Burbank (1933, p. 279–281), three principal geanticlinal areas, each trending from northwest to southeast across Colorado, completely altered the early Paleozoic pattern of seaways in the central part of that state, cutting off some of the seas from the Arizona area and at the same time furnishing a considerable amount of sediment that was transported southwestward into Arizona.

The oldest of the three geanticlinal developments, believed to have occurred in late Mississippian and early Pennsylvanian time, was located approximately where the Front Range is today and served to separate the sea into an eastern and a western tongue. It is described as sufficiently high to bring about "deep weathering and erosion" as illustrated in the sediments accumulated in adjoining areas.

A second area of uplift was to the west in the region of the Uncompahgre plateau and west of San Luis valley. Burbank dates these mountains as starting in early Pennsylvanian and continuing into Permian time. When they had developed, the Colorado seas were divided into three segments and although probably less than 100 miles wide, the mountains must have reached very great heights judging from the volume of detrital sediments deposited in the bordering areas. They have been called the Uncompahgre-San Luis highlands by Baker, Dane and Reeside (1933, p. 975). On the southwest flank, toward Arizona, was developed a deep structural basin in which much coarse debris and also the very thick saline deposits of the Paradox formation were accumulated during Pennsylvanian and Permian time.

The final uplift, in the San Juan upland still farther to the southwest, is described by Burbank as taking place in late Permian and Triassic time. It involved folding and faulting and influenced sedimentation in that area but was less significant than the other two.

Triassic-Jurassic Deposits

Pattern of Isopachs in Arizona

The pattern of positive elements and sinking basins that, with only minor modifications, characterized Arizona throughout the Paleozoic era experienced a radical change with the arrival of the Triassic (Pl. 2, C). Regional uplift in central and/or southern Arizona during this period resulted in sedimentation being confined to the northern half of the state. As in the earlier periods, the Cordilleran geosyncline existed to the northwest in the form of a constantly sinking trough and continued to influence deposition in Arizona through advances and retreats of its waters. Southwest of Arizona in Sonora, Mexico, the margins of a southern geosyncline which survived the Paleozoic are indicated by thick deposits of Triassic and Jurassic age. The Uncompahgre Ridge in Colorado persisted as a source of sediments and as a rim for the basin in the southwestern part of that state.

The isopach map of the Triassic period is based on thickness figures representing a combination of two distinct series of deposits formed under rather different environments. The older

of the two series (Moenkopi fm.) is of early Triassic age and forms a wedge that is 2000–3000 feet thick in western Arizona and Utah, but thins eastward until it ends along a line extending from western Colorado into eastern Arizona. Near the geosyncline on the western margin of this wedge, marine sediments form a large part of the deposit; farther east, continental redbeds occur alone. Deposits of late Triassic age (Shinarump, Chinle) overlie these older beds in Arizona and Utah, and extend far beyond their eastern limit into Colorado and New Mexico[4]. These Upper Triassic sediments are entirely continental and accumulated with centers of greatest thickness far eastward of those for the Lower Triassic. Thus combining the unlike patterns representing deposits of these two epochs gives a map with depression contours, showing relatively thin deposits, far westward of the basin's east margin.

Deposits of the Jurassic period (Glen Canyon group, San Rafael group, Morrison formation) form a pattern on the isopach map (Pl. 2, D) much like that of the Triassic. The southern margin of the Arizona basin is again near the central part of this state; contours indicating amount of sinking likewise increase north westward across southern Utah. As in the Triassic, evidence of a Mexican geosyncline is found in Jurassic deposits of Caborca, Sonora, not far south of the Arizona boundary. This depositional trough may have connected across southern California and southwestern Arizona with the Cordilleran geosyncline as shown by Eardley (1949, figs. 10, 12), but no evidence for this is known to the writer. Certainly no sediments of either Triassic or Jurassic age have yet been recognized in southwestern Arizona or adjacent parts of California.

In the basin of northern Arizona, southern Utah, and adjoining states, sedimentation apparently continued through late Jurassic time, as shown by the Morrison formation. Further south in Arizona, crustal disturbance and uplift apparently took place at this time, as illustrated by extensive gravel accumulations at the base of the Lower Cretaceous series in that area. This feature is discussed in the section on Lower Cretaceous deposits; suffice to say here that the disturbance in Arizona may have been related to the well-established late Jurassic Nevadan revolution of California.

Evidence of the Late Triassic Uplift

The first extensive uplift in Arizona after the Precambrian probably came during the Upper Triassic epoch. At that time the region south of the Colorado Plateau apparently was uplifted for the Shinarump conglomerate of Late Triassic age, deposited as a broad sheet of gravel over much of northeastern Arizona and adjoining areas, testifies to vigorous erosion with long transportation from an adjoining area. Among the gravels in this conglomerate are certain distinctive lithologic types and others containing diagnostic fossils that (McKee, 1937) prove an origin to the south for much of the sediment. An area of uplift must also have developed at this time in Colorado to the east; for, according to Lee (1926), a similar type of evidence occurs in that region.

Whether any of the Upper Triassic highlands were the result of squeezing and close-folding is not known, but apparently they were elevated sufficiently to allow a vigorous erosion, as shown by the large size and high degree of rounding of many of the gravels. Also, all but the most durable types of rock such as quartzite and chert are absent in the Shinarump conglomerate. Although the exact locality of the highlands is not known, a southern limit is determined by the presence of Upper Triassic marine deposits in northwest Sonora, Mexico (Cooper and Arellano, 1946, p. 611). Folding of a "Permo-Triassic" geosyncline in southwestern Arizona during Early Triassic time has been suggested by Tenney (1930, p. 275-277) in an hypothesis to explain the northwest trend of mountains in that part of the state; however, no evidence is given for the dating and no proof of early Mesozoic folding is cited.

Lower Cretaceous Deposits

Pattern of Isopachs in Arizona

A very complete change in the structural framework of Arizona developed at the start

[4] In southern New Mexico, Triassic strata probably are absent. According to H. Schmitt (Kelley, 1949, p. 28, 83), strata formerly considered Upper Triassic are to be correlated with the Abo formation of Permian age.

of the Lower Cretaceous period. Due to the Nevadan Revolution, the Cordilleran geosyncline, which previously had a tremendous influence on Arizona seaways to the east, was gone and former shelves and basins of deposition in northwestern Arizona were inactive. No sedimentation took place in the northern half of the state and only a thin sheet of deposits was laid down in Colorado and Utah to the northeast. On the other hand, a very considerable depression in southeastern Arizona brought about an invasion of the sea from Mexico (Pl. 3, A) and the accumulation of an exceptionally great thickness of sediments, both marine and continental.

The total area covered by Lower Cretaceous sediments in Arizona and adjoining portions of Mexico and the true thickness of these deposits in most parts of the region are very poorly known. The thickest section yet reported is in the Little Hatchet Mountains of southwestern New Mexico where Lasky (1947, p. 1) records over 21000 feet. A total of 16000 feet of supposed Lower Cretaceous strata was measured by Alexis (personal communication) in the Huachuca Mountains of extreme southern Arizona, and Brown (1939, p. 713–718) records over 8000 feet of sedimentary and volcanic rocks, determined by Reeside[5] to be probably Lower Cretaceous, in the Tucson Mountains. Rocks similar in lithologic character to those dated on the basis of fossils as Lower Cretaceous have been observed in many widely separated parts of southern Arizona. In most places they have been observed to have a considerable thickness, but measurements are difficult to make and are as yet not available. Outcrops in the Silver Bell and Vekol mountains and in various ranges westward across Yuma County in southwestern Arizona (Wilson, 1933) and northward at least as far as the New Water Mountains (McKee, 1947) are believed on lithologic grounds to belong to the Lower Cretaceous series.

Marine deposits of Lower Cretaceous age appear to be confined to the southeastern portion of the Arizona basin and probably do not extend much west of Bisbee (Ransome, 1904) and Patagonia Mountains (Schrader, 1915, p. 53). The extensive continental deposits over the rest of the basin contain plant remains, poorly preserved for the most part. In several localities, interbedded volcanic rocks form a considerable part of the series. Stoyanow (1949, p. 50–60) presents evidence to show that the configurations and connections of depositional basins changed several times, but he believes (p. 57) that available evidence "clearly indicates marine incursions from the south."

Deposits representing the Lower Cretaceous period have been shown (Reeside, 1944) to extend westward from the Rocky Mountain Geosyncline into northeastern New Mexico, across most of Colorado and across northern Utah. His isopach map indicates only moderate thicknesses of deposits over most of these areas and he states that sediments are largely nonmarine, but in northern Utah and adjoining areas, he shows a basin in which deposits are in excess of 3000 feet. Stokes (1944, p. 988) discusses the Lower Cretaceous deposits of western Colorado and eastern Utah, showing that in these areas they include a maximum of only a few hundred feet.

Evidence of Early Cretaceous (?) Uplift

The sedimentary record in southern Arizona gives clear evidence of widespread crustal disturbance and uplift between the end of the Paleozoic and the early stages of the Cretaceous. Extensive, thick conglomerate deposits in which gravels locally are of boulder size occur at the base of Cretaceous strata and above Paleozoic rocks in many localities. It is not certain that all such conglomerates are of the same age, for although in some areas the overlying sedimentary sequences are known to be of Early Cretaceous age, in others the dating is open to question and the deposits may be of Late Cretaceous age.

Best known of the late Mesozoic conglomerates of southern Arizona is the Glance conglomerate in the vicinity of Bisbee. It is the basal unit of a conformable sequence of rocks, determined on paleontologic evidence to be of Early Cretaceous age. According to Ransome (1904), the thickness of the Glance conglomerate north and east of Bisbee is, in general, between 50 and 75 feet, although locally the formation thins out entirely and in other places,

[5] Reeside, John B., Jr., in letter to J. H. Feth concerning identification of fossils recently collected in Tucson Mts., June 27, 1949.

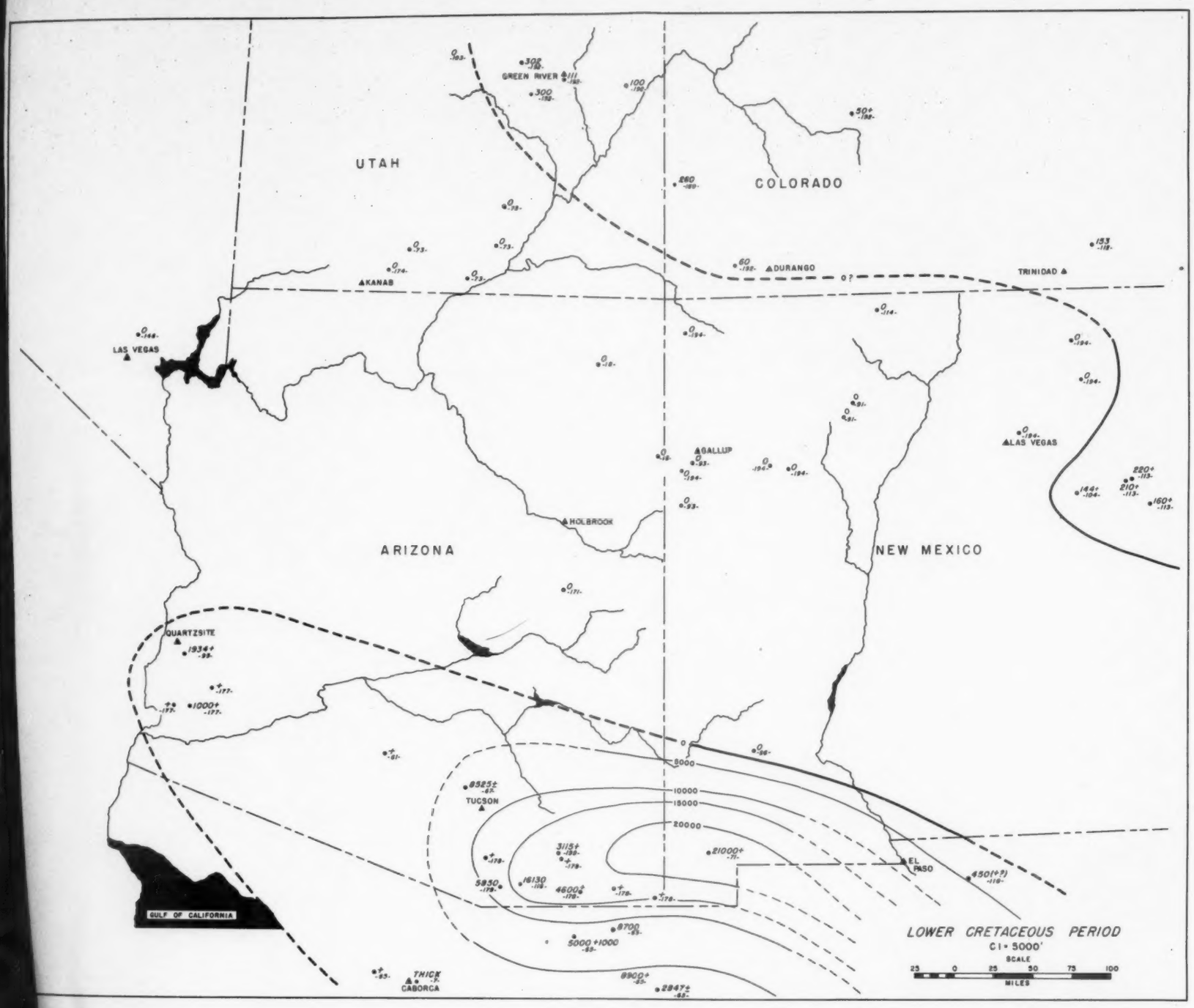
where it fills hollows in Carboniferous limestone, it has a total thickness of as much as 650 feet. Gravels in this conglomerate appear to be locally derived, with schists and limestones predominating. Granite, chert, and quartz pebbles, an admixture of sand and some red sandstone lenses are also present. Upward gradation is into dull-red or yellowish-brown shales.

Other Lower Cretaceous (?) conglomerates in southern Arizona have not been studied in detail, but some significant data on distribution and general character are available. On the east flank of the Empire Mountains a conglomerate, resting unconformably on Permian limestone and grading upward into red mudstone, is considered (Darton, 1925, p. 138–139) to be Lower Cretaceous because of lithologic similarity between the sequence to which it belongs and that of the Glance conglomerate at Bisbee. Brown (1939, p. 720) suggests Upper Cretaceous age for these beds because of a resemblance to strata thought to be of that age in the Tucson Mountains, but fossils recently collected in the Tucson Mountain deposits by Feth have been identified by Reeside (letter to J. H. Feth, June 27, 1949) as probably Lower Cretaceous. The conglomerate in the Empire Mountains contains gravels derived from various Paleozoic formations including those of Cambrian, Devonian, Mississippian, Pennsylvanian, and Permian age. Some are as much as a foot in diameter.

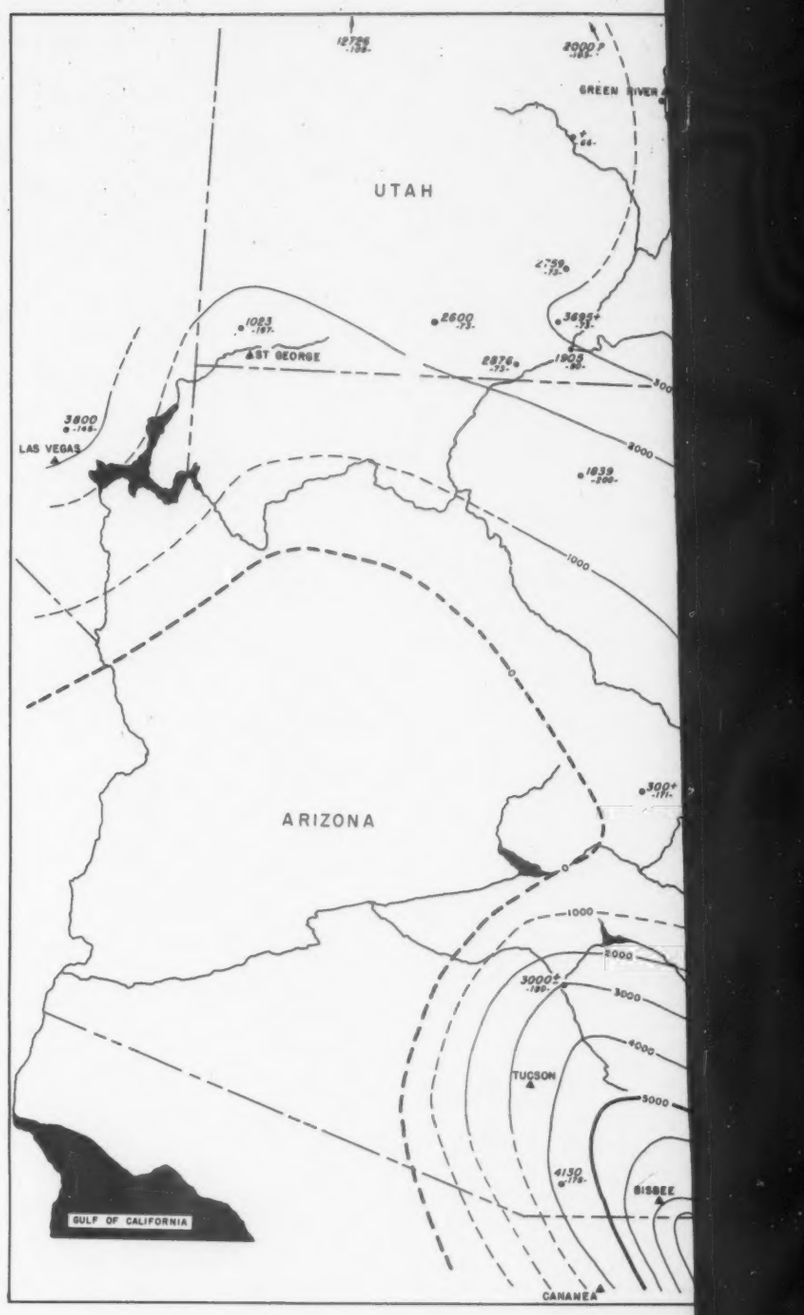
Basal conglomerates similar in character to the Glance are reported from near Helvetia in the Santa Rita Mountains (Darton, 1925, p. 139) and from the vicinity of Tombstone (Butler, Wilson, and Rasor, 1938, p. 23). B. S. Butler states (personal communication) that they are also found at Box Canyon gap in the Santa Ritas, on the east side of the Santa Catalina Mountains, in the foothills of the Rincon Mountains and high in the Huachuca Mountains. In none of these localities has the age been definitely determined on the basis of fossils. Except in the Huachuca Mountains where gravels in the conglomerate range up to the size of boulders several feet in diameter and are almost exclusively derived from Permian limestones, the conglomerates are characterized by small gravels and by the wide number of formations represented in the gravels. Furthermore, they are known to have been deposited on surfaces of considerable relief; in various localities they rest upon Cambrian quartzite, Escabrosa limestone, Naco limestone, and Permian limestone. Thus, the region must have been raised sufficiently to permit considerable and rapid erosion, to form valleys and canyons exposing rocks of many strata, and to allow the covering of this surface of relief with a widespread mantle of gravel, locally derived.

In western and southwestern Arizona, conglomerates that likewise may represent the base of the Cretaceous sequence and therefore may have resulted from the same crustal disturbances that caused the Glance, have been reported. These conglomerates include the locally derived boulder beds of the New Water Mountains described by the writer (McKee, 1947), the deposits of the Bill Baer Hills referred to by Stoyanow (1942, p. 1278), and beds in the vicinity of Growler Pass (B. S. Butler, personal communication). This apparent wide distribution of similar types of conglomerate, of local derivation and at the base of a sequence of Cretaceous strata, strongly suggests uplift of regional extent.

Because no evidence is known of folding during this uplift and because the high angle Dividend fault at Bisbee cuts Paleozoic but not Cretaceous strata (Ransome, 1904, p. 92), there is a strong suggestion that the uplift responsible for the conglomerates was largely achieved in the form of warping and/or block faulting. It appears likely from these data, moreover, that the conglomerates are all accountable to one general period of uplift and that the most probable date was early in the Cretaceous or approximately at the time of the Nevadan Revolution. An alternative hypothesis is that this uplift in southern Arizona was the same as the one that furnished the gravels of the Shinarump conglomerate during late Triassic time, but it seems scarcely probable that an uplifted area developed during that period should have remained long enough to influence sedimentation to the extent cited.



A LOWER CRETACEOUS



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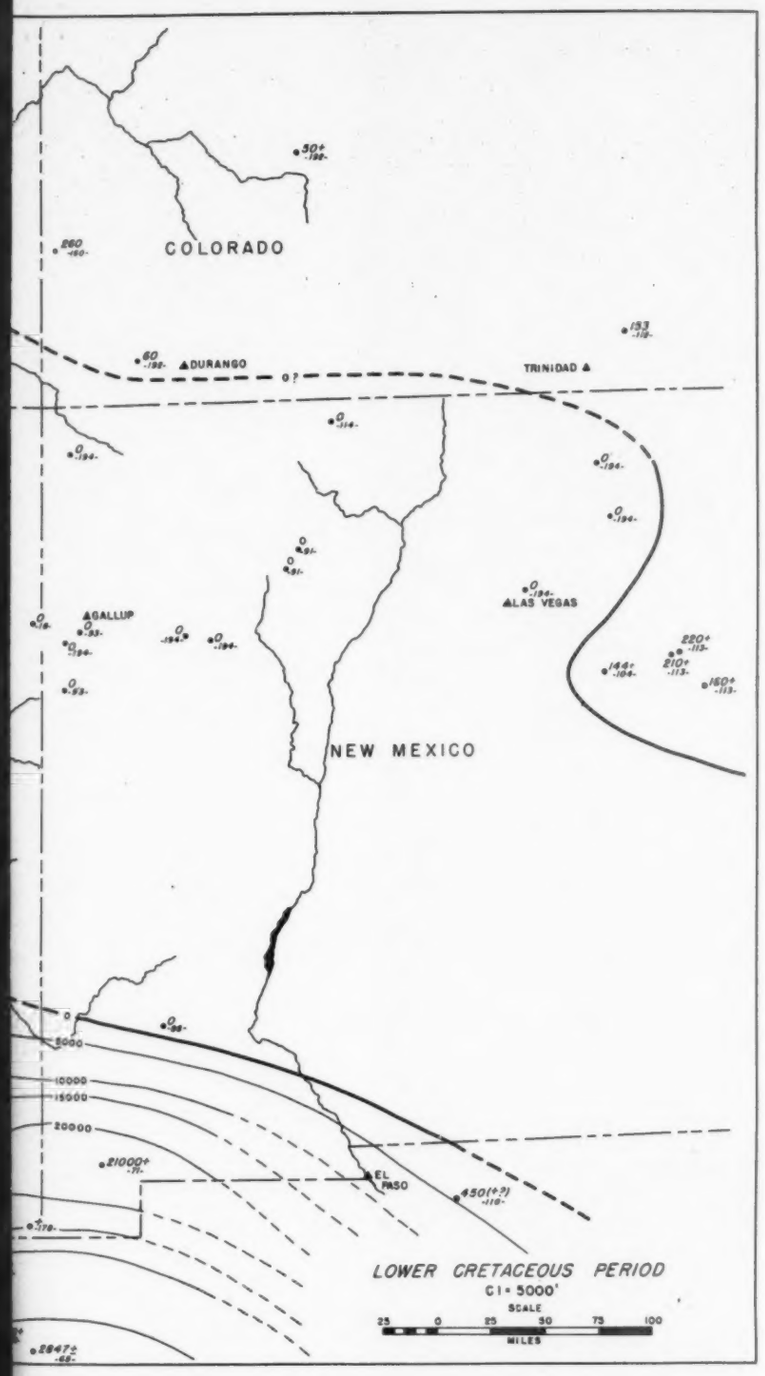
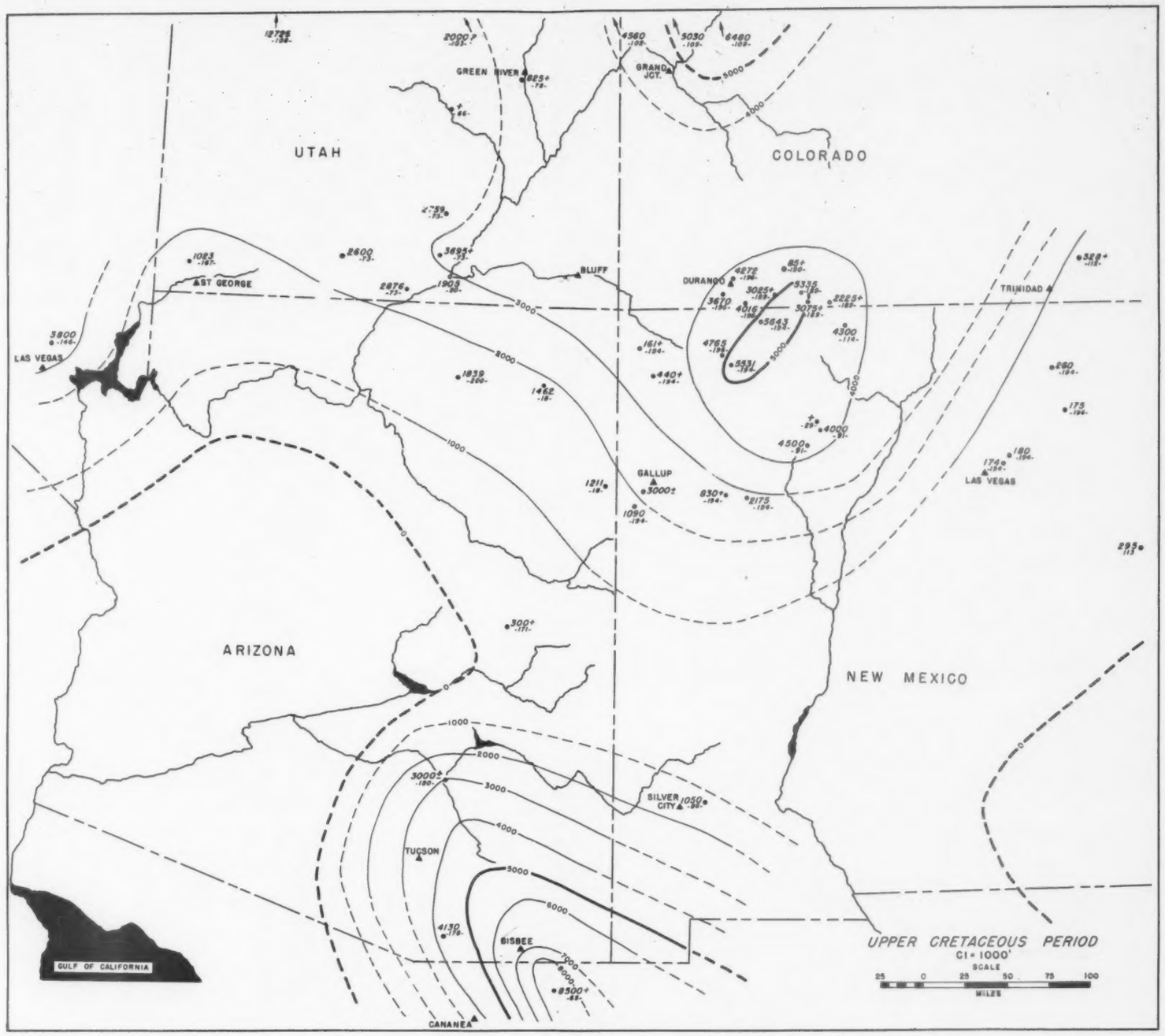
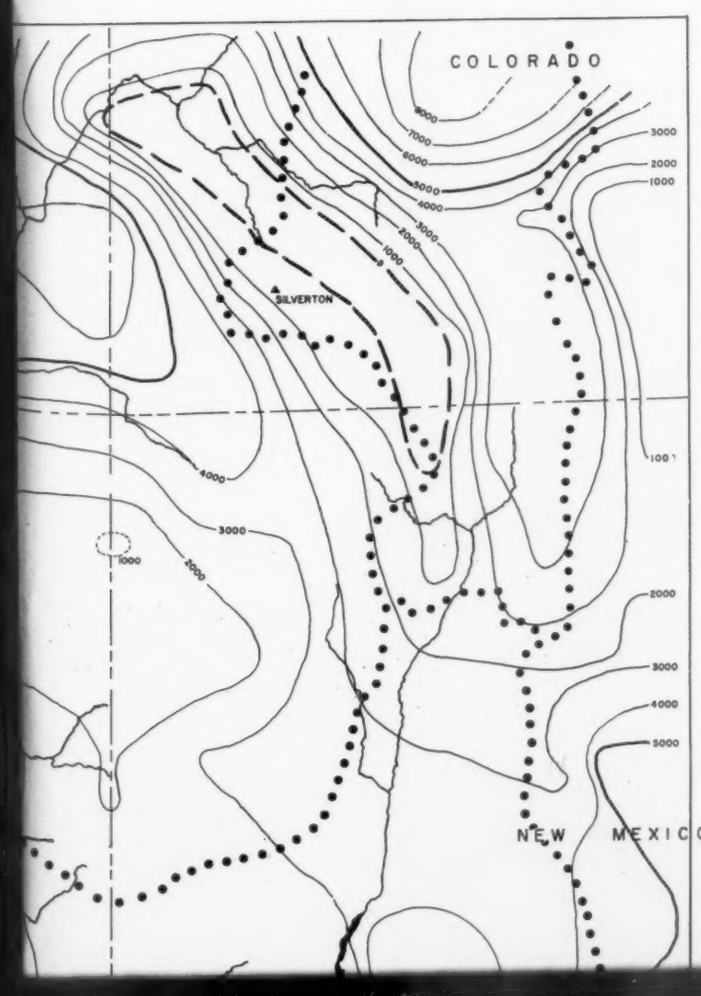
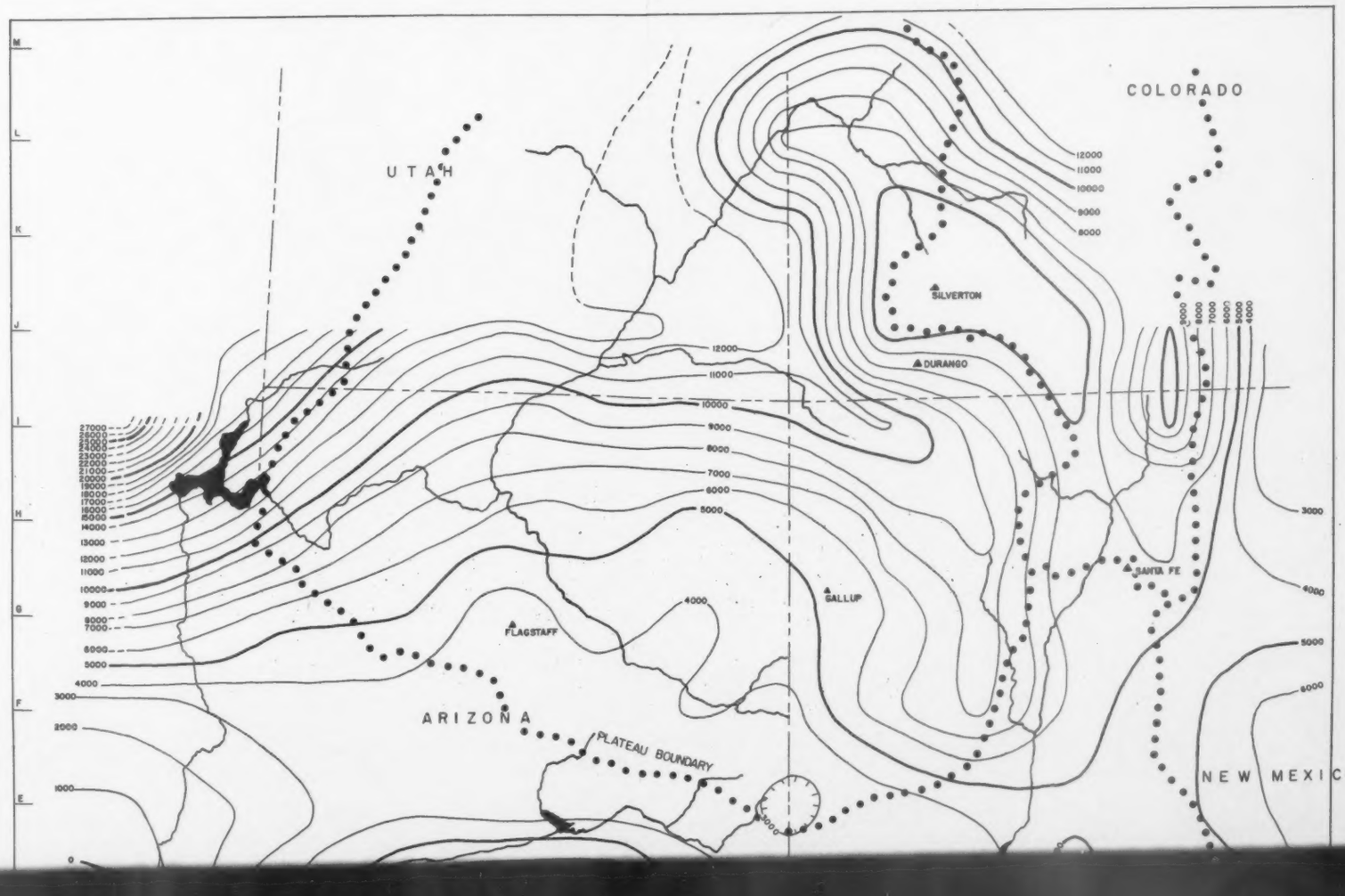
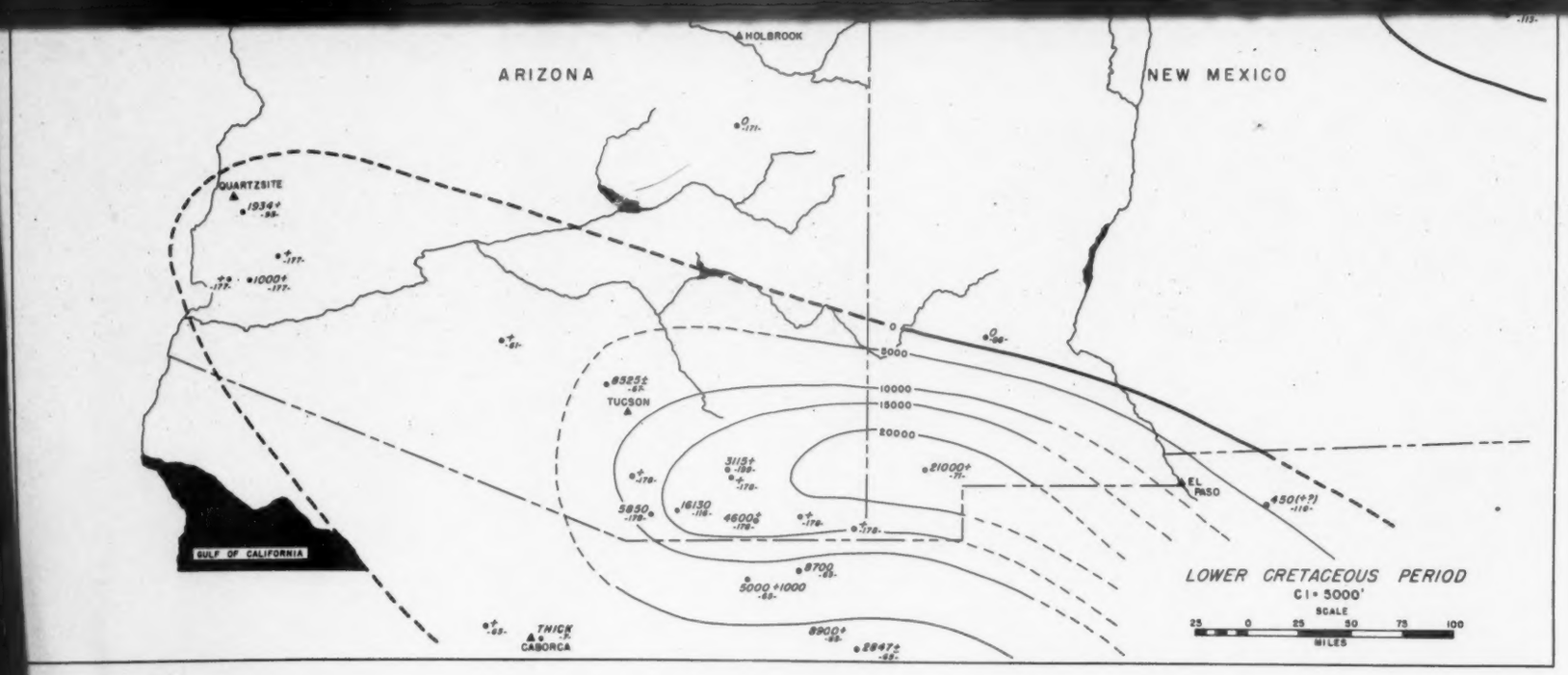




B UPPER CRETACEOUS







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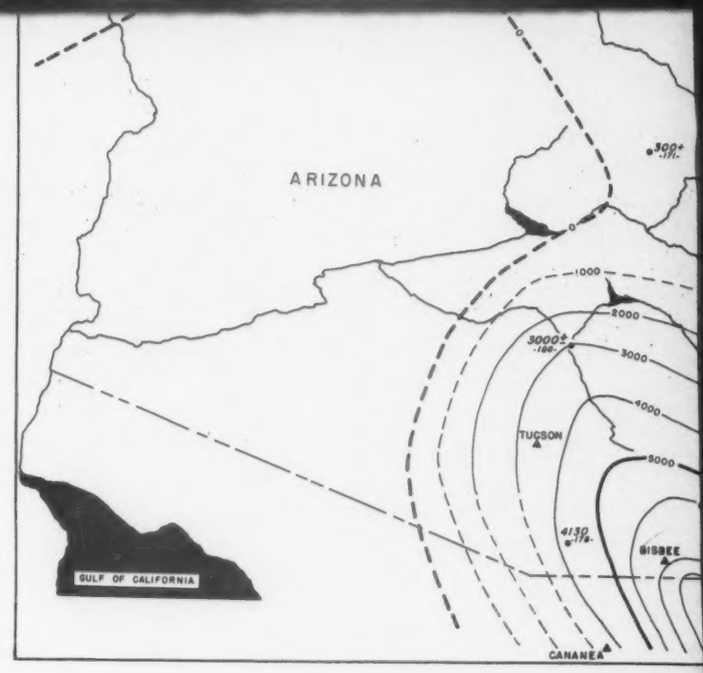

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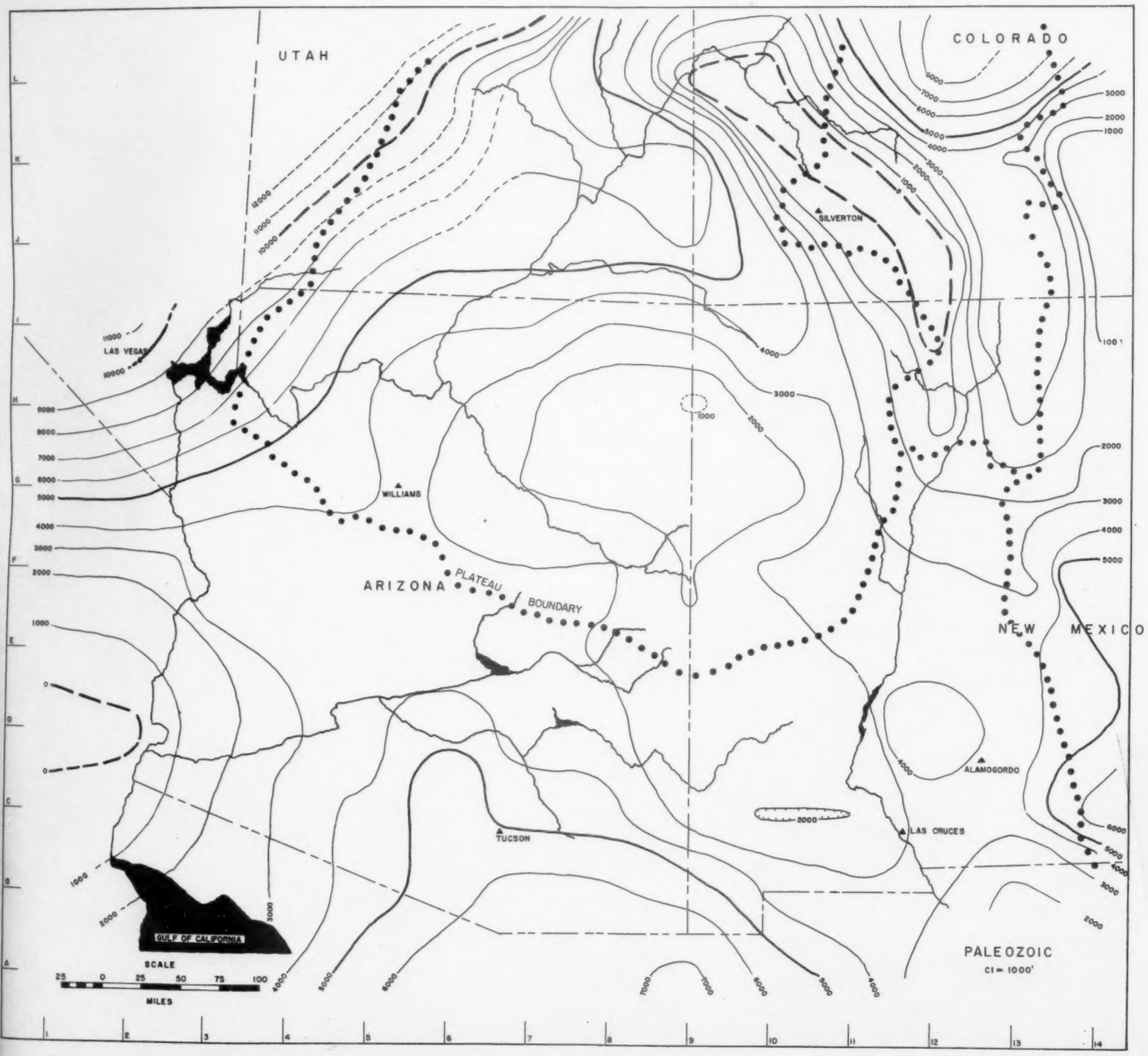

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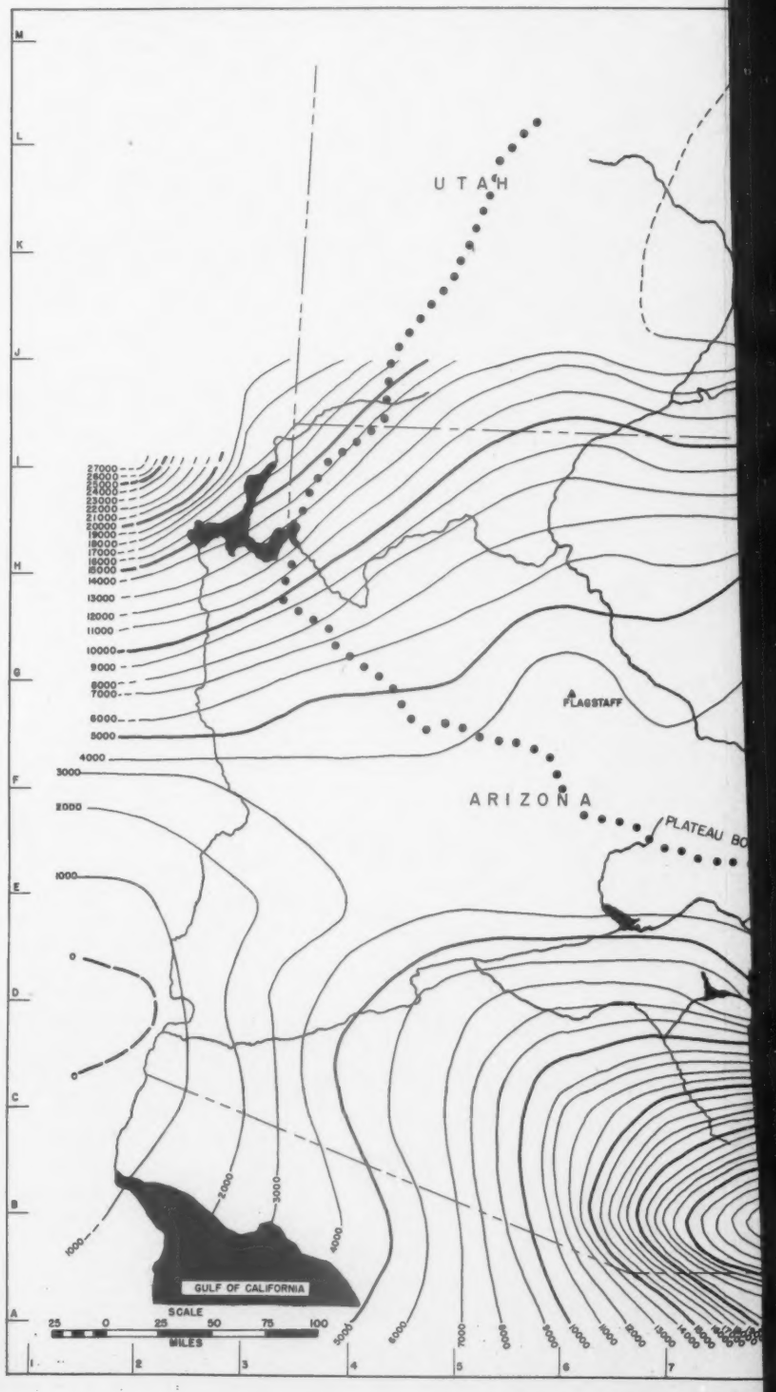

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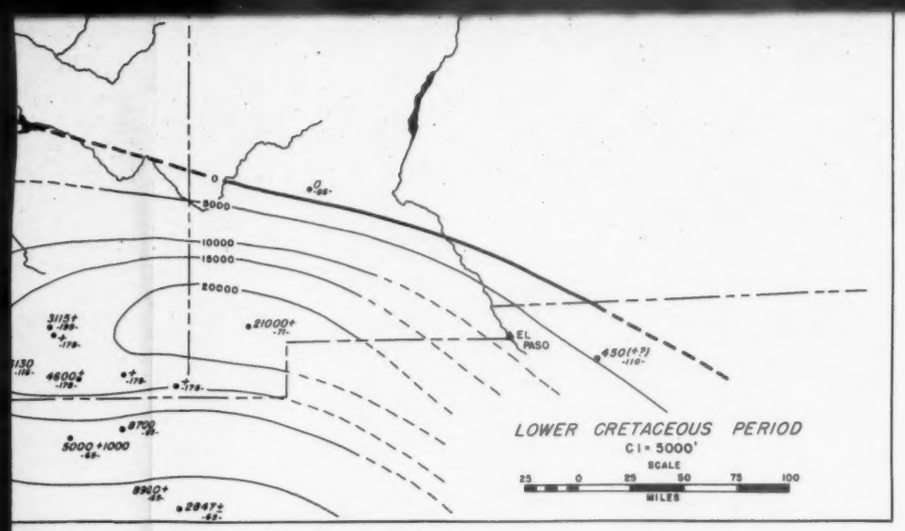

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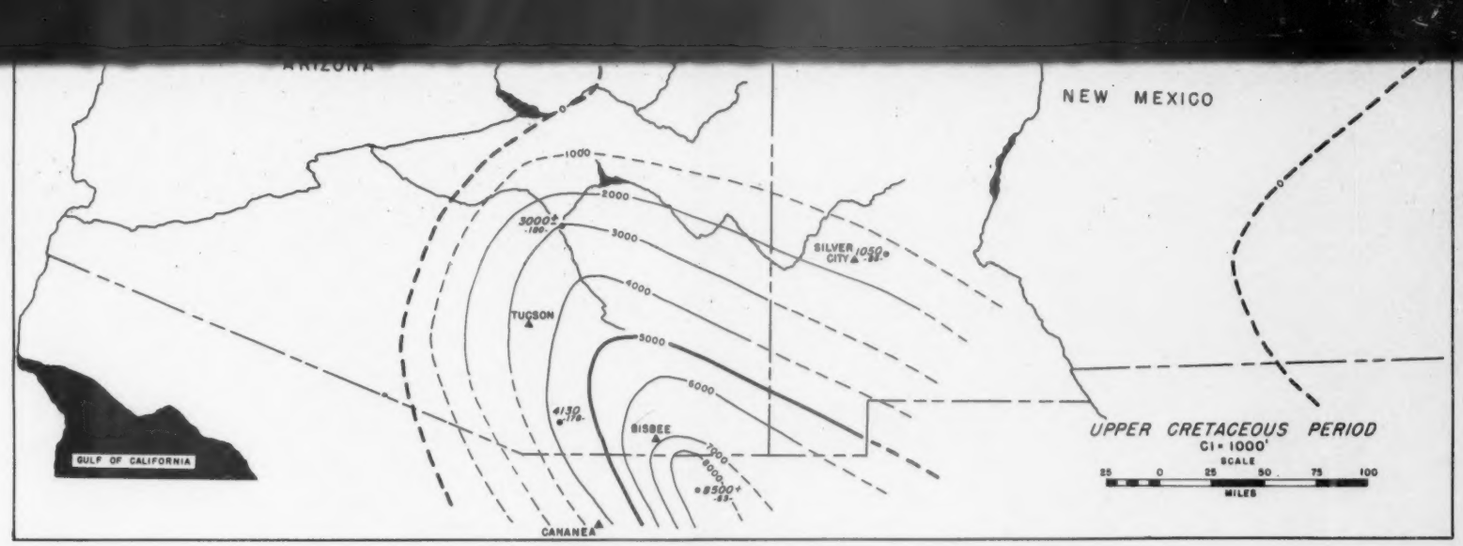

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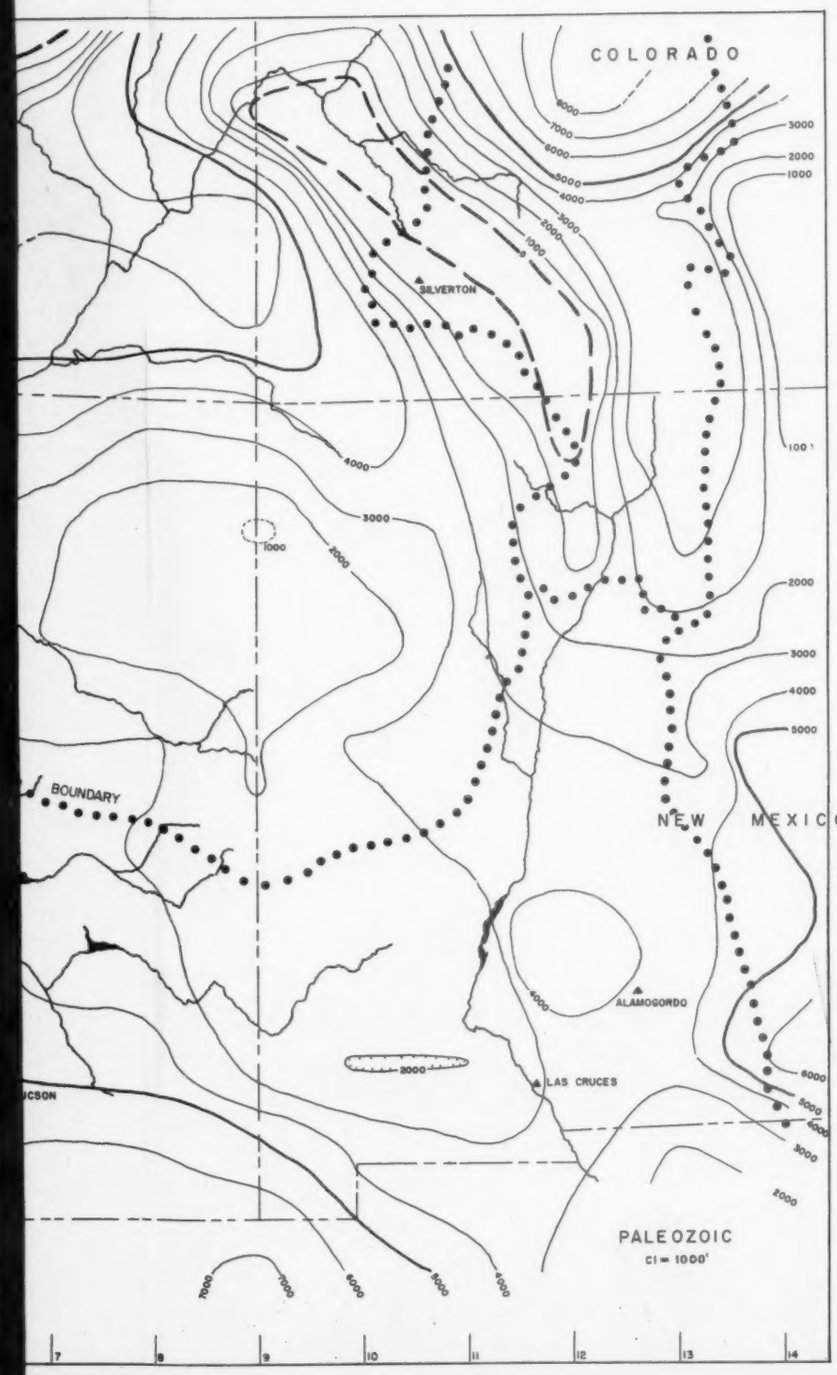

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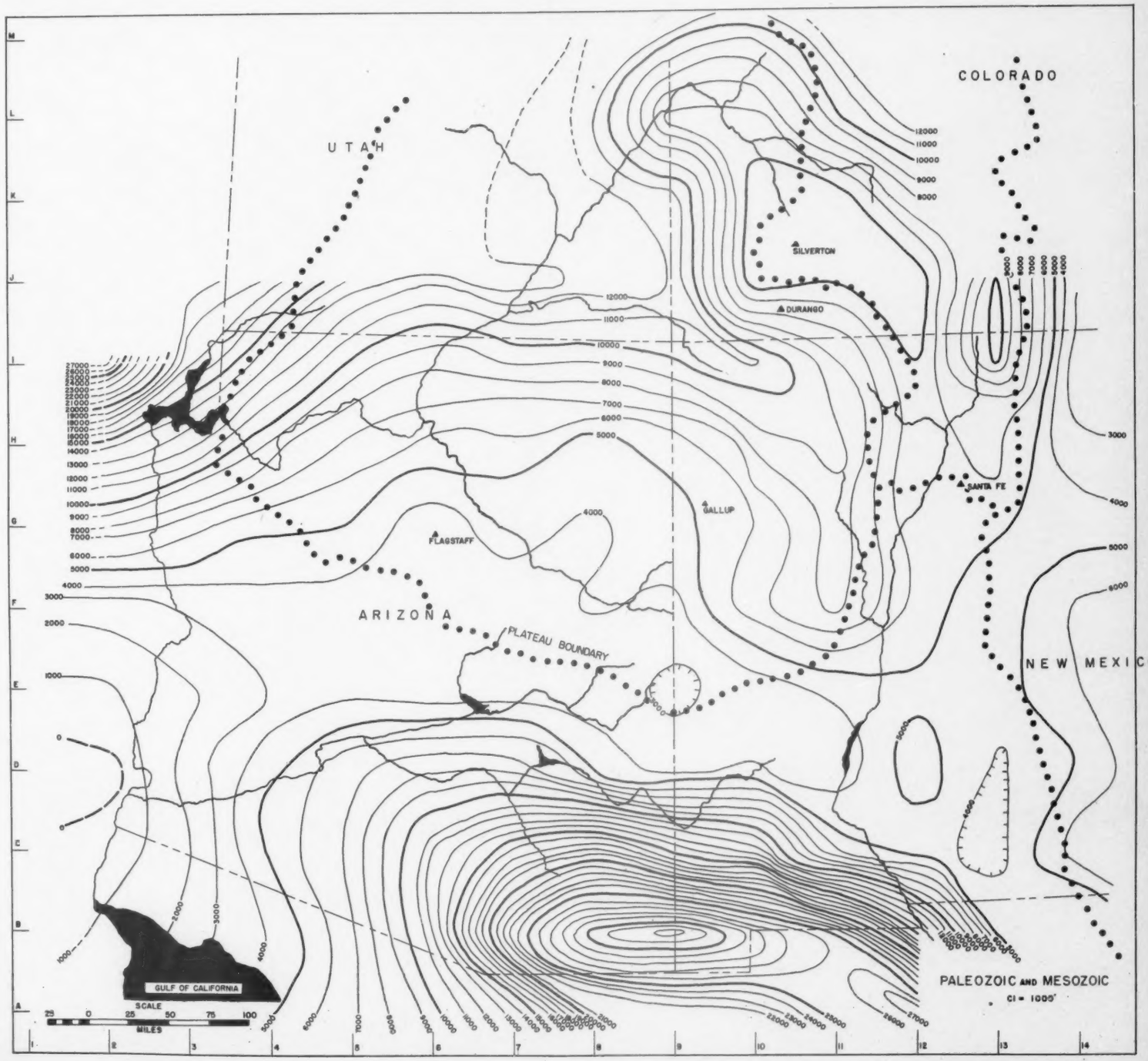

D COMBINED PALEOZOIC AND MESOZOIC

3. ISOPACH MAPS

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Upper Cretaceous Deposits

Pattern of Isopachs in Arizona

Thicknesses of Upper Cretaceous strata (Pl. 3, B) are difficult to represent effectively on an isopach map because of insufficient data. Because these deposits are younger and therefore are exposed over wider areas than are the underlying Paleozoic and Mesozoic formations, they have been affected by erosion far more extensively and their original thicknesses can not be measured in most areas.

Across the northeastern part of Arizona and southward as far as the Mogollon Rim, remnants of a sheet of Upper Cretaceous strata are found. Probably these strata once extended far westward of the present limit of outcrop in Arizona, for they are distributed with considerable thickness across the entire southern part of Utah and into Nevada. Probably, also, the sections in Arizona were once thicker than today for an unknown amount of sediment has everywhere been stripped from the surface.

Upper Cretaceous deposits of northern Arizona were formed during westward advances and subsequent withdrawals of the sea that occupied the Rocky Mountain geosyncline. Reeside (1944, map 2) has shown that although deposits were only a few thousand feet thick or less over most of the Rocky Mountain area, basins of rapid sinking in which 18,000 and 19,000 feet of sediments accumulated, were located in central and northeastern Utah.

In southern Arizona a deep basin of sedimentation appears to have developed during the Upper Cretaceous period, though little is known of the character and distribution of its sediments. Deposits of this age have been called the Sonoita group by Stoyanow (1949, p. 59) who has examined them on the east side of the Santa Rita Mountains. They have also been reported from the Christmas region farther north where they include andesitic lavas and pyroclastic rocks (Ross, 1925a p. 11–14; Ross, 1925b, p. 25–28) and in the northeastern part of Sonora, Mexico, to the south (Taliaferro, 1933). In this area evidence of volcanism is also found.

Dating of the various Upper Cretaceous sections in southern Arizona and Sonora has been partly on the basis of plant remains and partly from dinosaur bones. These indicate a late stage of the period as shown by Reeside (1944), in contrast to the early part represented by outcrops in northeastern Arizona.

Total Deposits of Paleozoic and of Combined Paleozoic and Mesozoic

From data compiled on isopach maps of the various Paleozoic and Mesozoic periods, composite maps showing total deposits in Arizona and adjoining areas have been prepared. The method used was to superimpose on all period maps a grid pattern divided each way into 13 units or squares; to determine for each square the average thickness of deposits of the period involved; and from these figures to calculate the total average thickness of strata in each square. These data were used in preparing two composite maps illustrating: (1) total Paleozoic deposits and (2) total deposits of Paleozoic and Mesozoic combined.

The composite map of the Paleozoic era (Pl. 3,C) emphasizes the prominence of certain positive and negative elements within the region. The Cordilleran geosyncline is responsible for a total thickness of 8000 feet of Paleozoic strata in the northwest corner of Arizona; the Sonoran geosyncline for over 6000 feet in the opposite corner. Both thicken considerably away from the state boundary in Nevada and Sonora, respectively. In contrast are the Defiance Positive Area in northeastern Arizona with less than 2000 feet, and the Ensenada Positive Area in the southwest with less than 1000 feet of sediment. The sag or strait in central Arizona connecting the two geosynclines has over 3000 feet of sediment throughout. Outside Arizona but closely adjacent in Colorado is the Uncompahgre-San Luis ridge containing no Paleozoic deposits, but with 8000 feet of sediments to the northeast and over 6000 in the Utah basin to the southwest.

The most striking features of the isopach map combining thicknesses of Paleozoic and Mesozoic deposits (pl. 3, D) are the extremely great accumulations in the northwest (over 25,000 feet north of Lake Mead) and the southeast (30,000 feet) corners of Arizona. Likewise

notable is a belt of sediments 12,000 feet thick across southern Utah and another 10,000 feet thick in north-central New Mexico, which contrast with the belt of relatively thin deposits (mostly under 4000 feet) that extends from west to east across central Arizona and New Mexico.

Cenozoic Deposits

Sedimentary Record of Late Miocene or Early Pliocene Uplift

A considerable amount of evidence accumulated within recent years indicates that crustal disturbance and uplift affected most parts of Arizona during late Miocene or early Pliocene time. As a result, thick deposits of sediment, consisting mostly of coarse detrital material, accumulated in many valleys and basins.

In southern and southeastern Arizona, strata of Pliocene age have been referred to the Gila conglomerate. First described by Gilbert (1875, p. 540–541) from the Gila River valley west of Clifton, similar strata of approximately the same age, shown through subsequent studies in San Pedro, Sulphur Spring, and other intermont valleys, probably represent a common period of accumulation. These deposits, within the Gila and San Simon valleys at least, are composed of two very distinct facies according to Knechtel (1936, p. 85). One is a fanglomerate that borders the mountain ranges. The other is composed of lake deposits including sandstone, tuff, limestone, and diatomite that underlie the valley centers and interfinger with the surrounding conglomerates.

The Gila conglomerate is dated as Upper Pliocene on the basis of vertebrate fossils. Collections from the San Pedro valley have been described by Gidley (1922; 1926); others from the Gila and San Simon valleys by Gazin (*in* Knechtel, 1936, p. 86–87). From evidence of pre-Gila deformation, shown by unconformity beneath the conglomerate and by steeply upturned Miocene (?) volcanics below, Knechtel (1936, p. 89–90) infers "that the latest deformation of mountain-building proportions that affected this region occurred prior to the deposition of the Gila conglomerate, in late Miocene or early Pliocene time."

In central Arizona, near Walnut Grove south of Prescott, evidence of extensive uplift during late Miocene or Pliocene time, resulting in thick deposits of both coarse and fine detrital sediments, has recently been discovered. The precise age of the deposits is still in question pending examination of extensive collections of vertebrate remains made during the summer of 1949 by a joint expedition of the Museum of Northern Arizona and the University of Arizona, but it is clearly of Pliocene time. Furthermore, interbedded lavas show that not only uplift sufficient to develop extensive fans but also volcanism must have been widespread in central Arizona.

In northern Arizona, along the southern borders of the Colorado Plateau, abundant scattered remnants of once extensive gravel deposits are believed to have developed in Pliocene time as a result of late Miocene or early Pliocene uplift to the south. These gravels occur eastward at least as far as Fort Apache, where they cover the tops of many buttes and mesas. Further west, they occur along the rims of Sycamore Canyon beneath basalt flows of supposed late Pliocene age (Price, 1950). To the north and west, other scattered outcrops are at Anita, Redlands, Rose Well, Frazier's Well, and several other localities. The deposits at Rose Well and Frazier's Well, which are several hundred feet thick, have recently been described by Koons (1948) and given the name of Blue Mountain gravels. Locally they may be seen beneath early basalts of that area.

Evidence that the Pliocene (?) gravels of the Colorado Plateau were derived from uplifted areas to the south has been discussed under "Former Southward Extension of the Plateau." The evidence consists primarily of the fact that these gravels have been derived largely from rocks of early Paleozoic and Precambrian age that could only have come from the south in order to reach their present resting places on eroded surfaces of Permian strata. Further evidence is found in the buried profiles of former steep-walled valleys as reported by Brainerd Mears (personal communication) in the upper walls of Oak Creek Canyon beneath basalt flows. These canyons appear to be relics of a former north and northeast-trending drainage system. The size of some of the boulders, notably those several feet in diameter of granitic

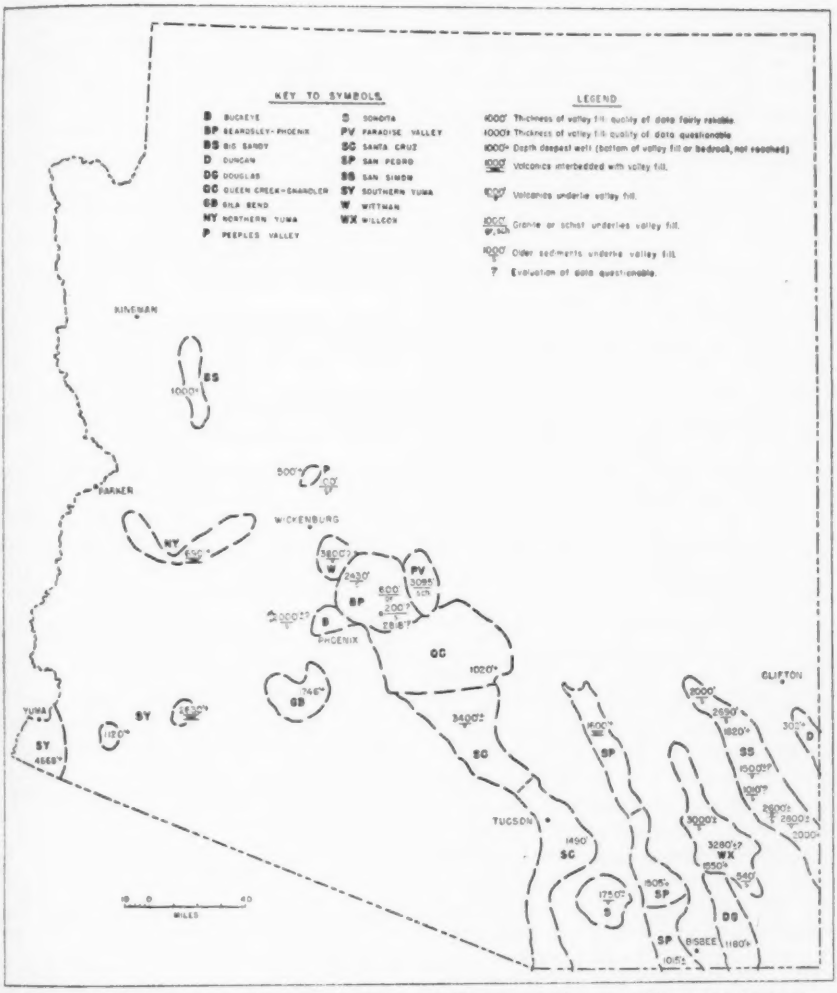

FIGURE 3.—LOCATION AND THICKNESS OF CENOZOIC DEPOSITS IN PRINCIPAL BASINS OF ARIZONA

rock at Rose Well, testify to the proximity of the uplifted source rock.

No fossils have yet been found among the Pliocene (?) gravels of the Colorado Plateau, so their age must be considered tentative. They are older than the earliest lavas of the San Franciscan field which have been placed in late Pliocene by Robinson (1913), Childs (1948, p. 379), and others. The gravels may therefore represent the results of rapid erosion caused by major uplift of the Colorado Plateau blocks, an event dated by Longwell (1946, p. 834) as late Miocene or early Pliocene. With the Prescott block raised higher than the blocks to the north, a north-flowing drainage would have been the normal result and adjustments to the present west-flowing Colorado River drainage in Grand Canyon and the south-flowing Verde River drainage would have followed.

Northeast of the Plateau margin, in the Hopi Butte Country, lake deposits known as the Bidahochi formation, of Middle or Late Plio-

cene age, have been described by Williams (1936) and Hack (1942). They are composed of sand, clay, water-laid marl, and tuff, up to 400 feet thick, containing fossil mammals, birds, and fish. They rest on an erosion surface of low relief. On stratigraphic grounds these beds have been shown to be older than the oldest lavas of the San Franciscan field so they may be roughly equivalent in age to the Pliocene (?) gravels farther south, previously described. If this is true, distance from the central Arizona upland and the slight relief then in the Hopi Butte area would account for the contrast between the fine sediments there and the coarse gravels that filled canyons and spread over the mountain base to the south. The surface on which this depositon occurred would correspond to the "regional Miocene surface of erosion" postulated by Gregory (1947, p. 704) as that "on which after uplift the Colorado River began its life as the master stream in an integrated, through-flowing, generally superposed, drainage system."

West of the Colorado Plateau in the area of Lake Mead are several thick series of Cenozoic deposits that have been described and discussed by Longwell (1946). They include the Muddy Creek formation of Miocene (?) age, composed of interior basin fanglomerates interfingering into fine clastics and salines, and later deposits composed both of river gravels and of lake beds (Chemehuevis formation). Above the Muddy Creek formation and also within the later gravels are angular unconformities testifying to the occurrence of crustal disturbances. The thickness of the various coarse detrital sediments indicates surfaces of marked relief at various times. Most significant, however, is evidence that a through-flowing Colorado River had its beginnings after deposition of the Muddy Creek formation, as shown by the general character and the widespread imbrication of the overlying gravels. Thus (Longwell, 1946, p. 834), although the exact age of the critical Muddy Creek formation is not yet fixed, fossil and other evidence favor an Upper Miocene age, which leads to the conclusion that the Colorado River began in its present course not earlier than Pliocene. This is the tentative date, therefore, when uplift of the Colorado Plateau blocks began.

Deposits of Late Cenozoic Time

In most parts of Arizona, it is not yet possible to separate strata of Pliocene, Pleistocene, and Recent age with any degree of certainty, though great accumulations of the combined deposits fill most of the valleys within the southern and western parts of the state. Little is known of the total depths of these valleys, but data from deep wells and, in some places, from geophysical probes show minimum depths of 2000 feet or more in numerous localities. Figure 3, prepared by S. F. Turner of the U. S. Geological Survey and L. A. Heindl of the State Land Department,[6] summarizes the data. Unfortunately few measurements give total depths to bedrock, nor do they give much idea of the shapes of buried valleys.

Crustal disturbances in Pliocene time probably inaugurated the great accumulation of detritus that fills most of the Arizona valleys, though an unknown amount of older gravels, faulted and deformed, may occur in some areas as suggested by the Whitetail conglomerate in southeastern Arizona and the Horse Spring and Muddy Creek formations of northwestern Arizona. Clearly, rejuvenation at various times following the initial uplift has controlled and increased the total accumulation within the various intermontane valleys. Information is not yet sufficient to permit any attempt to correlate such movements with the resulting sediments.

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FORM AND STRUCTURE OF SILLS NEAR PANDO, COLORADO

BY OGDEN TWETO

ABSTRACT

Four varieties of early Tertiary quartz monzonite and quartz latite porphyries: the Elk Mountain, Lincoln, and Eagle River porphyries, and one here named the Pando porphyry, form numerous sills in the Pennsylvanian and Permian(?) clastic rocks of the southern part of the Gore Range, Colorado. The major sills are 50–400 feet thick and 1–10 miles across. The sills fall into two groups on the basis of form and structure. Sills of the Pando and Elk Mountain porphyries, the more widespread and uniform although they show local irregularities in form and are locally discordant, are in general tabular in cross section and irregular in ground plan. Near the contacts, the porphyry of the sills shows folds, intrusive-stage faults, platy parting produced by differential movement between thin laminar plates of magma during intrusion, and linear mineral orientation at right angles to the orientation in the inner part of the sill and to the direction of magma flow.

Sills of the Lincoln and Eagle River porphyries are less extensive and more irregular in shape. They do not show intrusive-stage folds and faults except locally and poorly, and their linear mineral orientation is normally uniform from contact to contact and parallel to the direction of magma flow.

The structural features of the sills of Pando and Elk Mountain porphyry suggest high magmatic viscosity; absence of these structures in sills of Lincoln and Eagle River porphyries suggests that these were more fluid magmas, a conclusion that is supported by the difference in contact metamorphism produced by sills of the two groups. Contrary to what should be expected, the most uniform and widespread sills thus seem to have been formed by viscous magmas, and the most irregular sills, some of which grade into laccolithic bodies, by relatively fluid magma. Emplacement of the viscous Pando and Elk Mountain magmas in uniform and widespread sills may have been aided by the lifting action of laterally compressed competent beds. The Lincoln and Eagle River sill magmas apparently were capable of making room for themselves by lifting the load; in doing so they compacted the host rocks, which being lenticular yielded unevenly, causing smallscale knobs, domes, swellings and constrictions to form in the sills. Large-scale or laccolithic doming occurred near the supply vent, probably as a result of rapid introduction of magma.

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Introduction

The sills observed at several localities in Colorado show wide variations in form and structure. Although generally concordant and regionally blanket-like, they are locally discordant, and many are irregular in detail. Some sills are uniform in thickness over wide areas; some pinch and swell markedly; and some bulge into laccolithic bodies. The horizon of some sills is marked by intrusion breccia beyond the limits of the sill, and the horizon of others is an undisturbed bedding plane. Sills in certain types of clastic sedimentary rocks migrate extensively in stratigraphic position within short distances without developing notable discordance. Changes in form and horizon of some sills are caused by faulting in the wallrocks, but similar changes occur in other sills in the absence of wallrock faulting. Sills in thick, moderately deformed sedimentary sections differ in form and structure from sills near volcanic centers, and sills closely related to laccoliths differ in some respects from both.

Some sills show only the simple flow structure defined by mineral orientation (Balk, 1937, p. 7–25). Others have a complex structure that includes both complicated mineral-orientation structure and larger scale structures such as intrusive-stage folds, faults, and breccia. Folding caused by primary flow in intrusive rocks is itself unusual (Cloos, 1946, p. 45) and the mineral orientation structures seem more complex than those of most intrusive bodies because variations in flow structure are compressed into small areas. Markedly different orientations may exist within surface areas of only a few square feet, and in certain zones the two orientations commonly regarded as characteristic of the igneous and the metamorphic rocks may be present together.

This study is based principally upon observations in the southern part of the Gore Range, in Eagle and Summit counties, Colorado, east of Pando and north of Leadville (Fig. 1). Here, and in the Kokomo district to the east (Koschmann and Wells, 1946), the range is made up primarily of Paleozoic sedimentary rocks intruded by many sills of monzonitic porphyry. The Paleozoic rocks comprise about 500 feet of pre-Pennsylvanian quartzites and dolomites and almost 8,000 feet of Pennsylvanian and Permian (?) grits, shales and conglomerates of the Minturn and Maroon formations (Tweto, 1949). At least 30 major sills are present in the area, chiefly in the Minturn and Maroon formations. Porphyry occurs also in many minor sills, a few dikes and irregular bodies, and in one large laccolithic body, on the south side of the East Fork of Eagle River in an area as yet unmapped in detail. The sedimentary rocks and sills form a northeast-dipping homocline which is intensely faulted in the eastern part of the Pando area. A geologic map of the area prepared for the U. S. Geological Survey will presumably be available by the time this paper is printed (Tweto, Preliminary geologic map of the Pando area, Eagle and Summit Counties, Colorado).

A few of the observations were made at Green Mountain, an isolated mountain in Blue River Valley near the north end of the Gore Range, in Summit and Grand counties. Green Mountain presents a 2,000-foot section of trachytic sills in the Cretaceous shales. The sills are pierced by a probable volcanic neck, and the mountain appears to be a late Tertiary volcanic center.

Field Work and Acknowledgments

The present paper, a product of a general geologic study of the Holy Cross quadrangle,

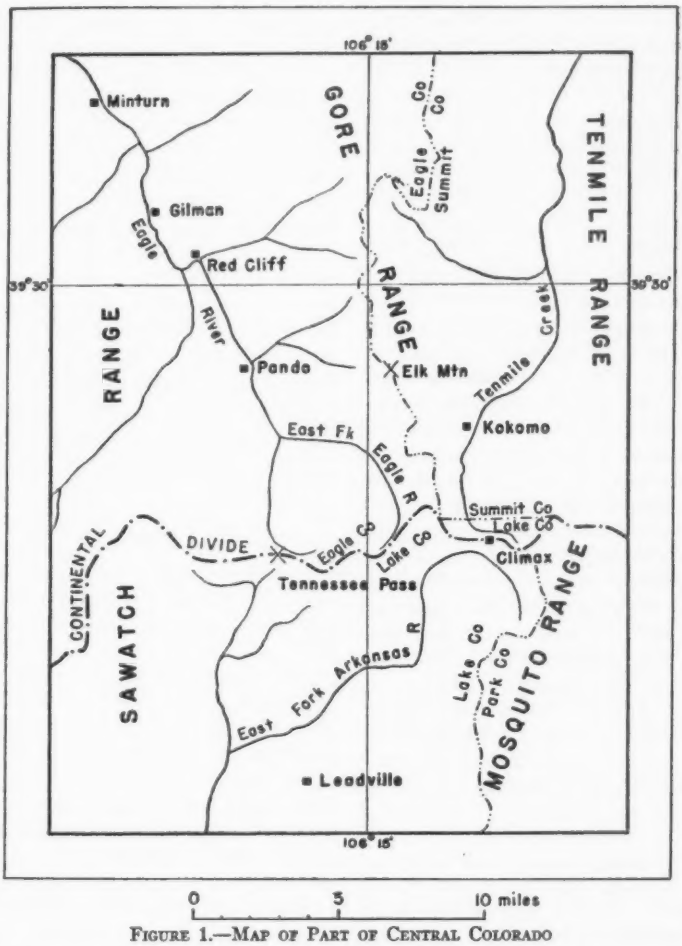

FIGURE 1.—MAP OF PART OF CENTRAL COLORADO

Colorado, for the U. S. Geological Survey in co-operation with the Colorado Geological Survey Board, is based principally on field work during 4-month field seasons in 1945 and 1947 and 1 month in 1948, and in part on work in other areas in central Colorado, particularly the Leadville, Green Mountain, and Gilman districts, at various times since 1940. Most of this work was done with the advice, guidance, or co-operation at various times of T. S. Lovering, W. S. Burbank, and A. H. Koschmann, to whom the writer expresses gratitude for critical reading of the manuscript as well as for other aid. Thanks are also due Dr. Earl Ingerson and Prof. Ernst Cloos for critical reading of the manuscript.

THE PORPHYRIES

General

The sills of the Pando area comprise four types of quartz monzonitic porphyries known, from oldest to youngest, as the Pando, Elk Mountain, Lincoln, and Eagle River porphyries, all of early Tertiary age. The Lincoln and Elk Mountain porphyries were first described by Emmons (1886, p. 78–79; 1898, p. 2–3) and

since then have become widely known in the Leadville region. The rock here named the Pando porphyry is quartz latite porphyry that forms a persistent sill or zone of sills in the Pennsylvanian shale just above the Leadville limestone and occurs also in small sills in the pre-Pennsylvanian quartzites and dolomites and in a few dikes. The main sill of Pando porphyry is persistent for several miles along the upper Eagle River, from the East Fork northward beyond Gilman, and crops out boldly on the slope just east of Pando, where it overlies a prominent cliff of silicified Leadville limestone. The Pando porphyry is the oldest of the four, as it is cut by the Elk Mountain porphyry, which is older than the Lincoln porphyry, and by the Eagle River porphyry.

The term Eagle River porphyry is here used for a quartz monzonite porphyry that is later than the Lincoln porphyry. Originally (Emmons, 1886, p. 88, 330–331), the term included both this later porphyry and the Lincoln porphyry. Emmons and Cross used the term only in the Leadville monograph (Emmons, 1886) and made no mention of it in their later report on the Tenmile district (Emmons, 1898, p. 2–3). The Eagle River porphyry is similar to the Johnson Gulch porphyry of the Gray porphyry group at Leadville (Emmons, Irving, and Loughlin, 1927, p. 48–50), and mapping now in progress may prove it to be equivalent.

Pando Porphyry

The Pando porphyry is a quartz latite porphyry composed of scattered quartz and plagioclase phenocrysts and ragged grains of altered biotite in a light-gray or buff, dense groundmass. (*See* also Crawford and Gibson, 1925, p. 30–31). The quartz grains are bipyramids that have been strongly rounded by resorption and are mostly about a fourth of an inch in diameter. The plagioclase crystals are thick tabular to prismatic and average about a fourth of an inch long. Biotite was once a constituent of the rock but has all been altered to pale-green chlorite or iron-stained sericite. Flow structure is defined almost entirely by the orientation of the plagioclase grains; the round quartz grains and ragged masses of chlorite give little or no suggestion of a preferred orientation. The plagioclase defines a rather weak linear orientation and a poor planar orientation, but the crystals are not abundant, and many of them are nearly equant, so that determination of structure from mineral orientation is usually difficult. A real orientation does exist, however; observations made independently by geologists who are experienced in determining orientations in porphyries give remarkably consistent results. Chilled Pando porphyry has a banded, glassy appearance, but the microscope shows that most of it is very finely crystalline. Phenocrysts of quartz and plagioclase are sparse but about the same size as those in unchilled porphyry. A weak linear orientation can be detected in most chilled Pando porphyry, but planar orientation is indiscernible.

Elk Mountain Porphyry

The Elk Mountain porphyry is a quartz monzonite porphyry consisting of abundant phenocrysts of quartz, plagioclase, and biotite in a gray, dense groundmass. The quartz grains are rounded bipyramids slightly modified by the prism and are a fourth to half an inch in diameter. Plagioclase forms moderately elongated tabular crystals a fourth to three-eighths of an inch long. Biotite is in hexagonal to subcircular plates typically about a tenth of an inch in diameter and in hexagonal books and prisms. Mineral parallelism is only weakly developed but with practice can be discerned in most outcrops; consistent results are obtained if several observations are made in the same locality, or if an observation at a given exposure is repeated at some later date. Mineral parallelism is defined chiefly by the orientations of biotite and feldspar grains; the subspherical quartz grains are of little determinative value except where locally concentrated in layers parallel to the sill contacts. Planar structure is outlined by biotite flakes and to a lesser degree by tabular plagioclase crystals; linear structure is defined chiefly by elongated plagioclase grains and to a minor extent by relatively rare biotite prisms and ellipsoidal quartz grains. In the chill zones, the feldspar and biotite phenocrysts are less than an eighth of an inch

in diameter, and the groundmass appears dark and glassy, although microscopically much of it is very finely crystalline. Quartz grains are the coarsest constituents of the chilled porphyry, and some are as large as the grains in the unchilled rock, although less abundant. Vague mineral orientation structure can be seen at places in the chill zones, but the structure is more easily determined in the unchilled porphyry with similar orientation adjacent to the chill zones.

Lincoln Porphyry

The Lincoln porphyry is a quartz monzonite porphyry similar to the Elk Mountain porphyry except that it contains large phenocrysts of light-gray or pink orthoclase. The orthoclase phenocrysts are typically about an inch long, but locally a few reach lengths of 2 or 3 inches. In a few sills, the orthoclase crystals are little if any coarser than those of plagioclase, and the Lincoln prophyry then closely resembles the Elk Mountain porphyry. Three varieties of the Lincoln porphyry are represented in the Pando area. Two varieties, one with very abundant orthoclase phenocrysts and one with very few and unevenly distributed phenocrysts, make up most of the sills and dikes of Lincoln porphyry in the area; the features described as typical of Lincoln porphyry sills in most of the discussion refer to these two varieties. The third variety is characterized by sparsely but evenly distributed phenocrysts of clear or glassy orthoclase. It forms a widespread sill that resembles sills of the Elk Mountain porphyry in certain features of form and structure.

Planar flow structure in the Lincoln porphyry is defined chiefly by the orientation of biotite flakes and, to a lesser extent, of the plagioclase and orthoclase crystals. Linear structure is defined primarily by the inequant plagioclase grains and a few of quartz. The prismatic orthoclase crystals show linear orientation at places, but locally they show no preferred orientation even though plagioclase and quartz do.

The Lincoln porphyry does not show as consistent nor generally as pronounced chilling as do the Elk Mountain and Pando porphyries. The chill zones are commonly a few inches thick, but they are absent in some places and as much as 4 feet thick in others. The chilled porphyry is fine-grained, but most of it lacks the dark, glassy appearance of the chilled Elk Mountain porphyry. It contains abundant but small phenocrysts of plagioclase, biotite, and quartz. Small grains of orthoclase are also present at most places, but the large orthoclase crystals characteristic of the Lincoln porphyry have been observed only locally in the chill zones of one sill.

Eagle River Porphyry

The Eagle River porphyry has a coarser groundmass and smaller phenocrysts than the Lincoln and Elk Mountain porphyries. It is a quartz monzonite porphyry composed of abundant phenocrysts of feldspars, quartz, and biotite in a gray or greenish crystalline matrix. The quartz is in rounded grains about a fourth of an inch in diameter and in slightly smaller rhombohedrons that have an almost square cross-section. Most of the feldspar grains are chunky and roughly equant, but a few of the larger glassy plagioclase grains are somewhat tabular. Biotite occurs in books that are almost as thick as they are broad and in flakes less than one-eighth inch in diameter. Owing to the paucity of inequant grains, mineral parallelism is poorly defined at best. A weak planar orientation defined by biotite flakes and a few tabular plagioclase grains can be discerned at most places, but linear orientation, which is defined by a few elongate feldspar and quartz grains, can be seen only locally and is invariably vague. The chill zones are narrow and are only slightly finer grained than the normal porphyry. They show a slight concentration of biotite.

General Relations of the Sills

The sills of the Pando area are unevenly distributed geographically and stratigraphically. Sills of the Pando, Lincoln, and Eagle River porphyries in general are concentrated in the southern part of the area, near the East Fork of Eagle River, and stratigraphically low in the section, near the base of the Pennsylvanian rocks. Sills of the Elk Mountain porphyry are most abundant in the eastern

part of the area and are concentrated higher in the stratigraphic section, lying for the most part above a horizon about 2000 feet above the base of the Pennsylvanian rocks. This distribution directly reflects the geographic and stratigraphic positions of the centers of intrusion. The Lincoln and Eagle River porphyry magmas apparently originated in a laccolithic body, or group of superimposed domed sills, on the south side of the East Fork of Eagle River, and they spread northward in the stratigraphically lowest shaly or thin-bedded zones, in the lower part of the Pennsylvanian section. The Pando porphyry magma also originated in this same area, or farther south, and spread northward. Sills of all three of these porphyries are most numerous and thickest near the East Fork of Eagle River, and they thin and die out to the north. The Eagle River porphyry is exposed for less than a mile north of the river but may continue farther northward beneath cover. Sills of Lincoln porphyry extend about 4 miles north of the river, and the main sill of Pando porphyry extends about 10 miles.

The Elk Mountain prophyry originated in the Kokomo district, east of the Pando area, probably at an intrusive center on Tucker Mountain, about 3 miles north of Kokomo (Koschmann and Wells, 1946, p. 74–77), not, as Emmons assumed (1898, p. 2), at Elk Mountain, on the divide between the Pando area and the Kokomo district. Detailed mapping of the east slope of the mountain by Koschmann and of the west slope by the writer has shown that the extensive bodies of Elk Mountain porphyry forming the mountain are thick sills, and that no cross-cutting bodies of any type are present there. The sedimentary section near Tucker Mountain is much thinner than in the Pando area, as all the sedimentary formations of the Pando-Kokomo region thin eastward by overlap against the ancient Paleozoic Front Range highland (Tweto, 1949, p. 155–158), and except for remnants of the Cambrian quartzite, the lowest beds are 3 to 4 thousand feet higher stratigraphically than those near the Eagle River. Thus sills of Elk Mountain porphyry spreading outward from the Tucker Mountain center had access only to the upper part of the Pennsylvanian-Permian(?) section and were concentrated there, but a few formed at somewhat lower horizons farther west, fed presumably by dikes such as one that cuts the Precambrian rocks across the river from Pando. The sills of Elk Mountain porphyry are the most numerous and, except for one sill of Pando porphyry, the most widespread in the Pando area. They spread northwestward, westward, and southwestward at least 6 miles from the Tucker Mountain center and probably spread southward even farther.

Sills of the two older porphyries, Pando and Elk Mountain, differ from those of the Lincoln and Eagle River porphyries in form and structure, and probably in mode of intrusion. For this reason, the sills are considered in two groups in much of the discussion that follows.

Form

General

By definition, sills are tabular concordant intrusive bodies, but within these limits an almost endless number of geometric forms can occur. There are many variations in over-all shape in both plan and section, and irregularities and peculiarities of the host rocks and their structure produce many kinds of local deviations from the ideal sill form. The shape of a sill, particularly in ground plan, is in many cases indeterminate, because only one eroded edge of the sill is exposed, but the data available suggest that few if any sills spread out evenly from a central source like a flat and symmetrical pancake. In the Henry Mountains, where the structure and good exposures allow observation of some intrusives almost in their entirety, Hunt (1946, p. 12) has recently found that many of the sills are tongue-shaped and are marked by an anticlinal structure in the overlying beds. The sills of the Pando area, and of central Colorado in general, are definitely not of this shape, although some have tongue-shaped protuberances as much as a mile or two long. An irregular domal rather than anticlinal structure of the overlying beds should be expected of the sills of the Pando area, but the many superimposed sills and their uneven horizontal distribution prevents recognition of any consistent structure attributable to in-

dividual sills except locally, as on the north side of the East Fork of Eagle River, where the sediments are domed above a group of sills that are thickened in this area.

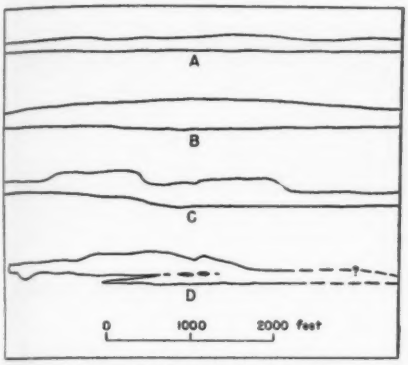

FIGURE 2.—PROFILES OF SILLS

Sketched from geologic map. A, typical tabular sill of Elk Mountain porphyry; B, typical lenticular sill of Lincoln porphyry; C, flat-topped domes on sill of Lincoln porphyry; D, typical irregular sill of Eagle River porphyry.

Shape

Sills of the Pando and Elk Mountain porphyries are on the whole tabular, with either blunt or slender wedge-shaped edges. Except near the edges, they show little or no progressive change in thickness (Fig. 2-A). They do show local variations in thickness. A few of the sills of Elk Mountain porphyry locally bulge gradually to as much as twice the normal thickness of the individual sills, but, compared to the horizontal extent, the changes in thickness are almost negligible and have little effect on the over-all shape. The Elk Mountain porphyry sills range from 5 or 10 to more than 300 feet thick, but most of them are 50 to 100 feet thick. Sills of the Pando porphyry range from a few inches to about 100 feet thick; the main sill is 50 to 75 feet thick through most of its area of exposure.

Sills of the Lincoln and Eagle River porphyries are less uniform than those of Pando and Elk Mountain porphyries and in general show lenticular rather than tabular forms (Fig. 2-B). All observed thus far thicken markedly in the source direction, and some are continuous with the laccolithic body on the south side of the East Fork of Eagle River. The sills also characteristically pinch and swell abruptly. In the Lincoln porphyry sills, the swells form

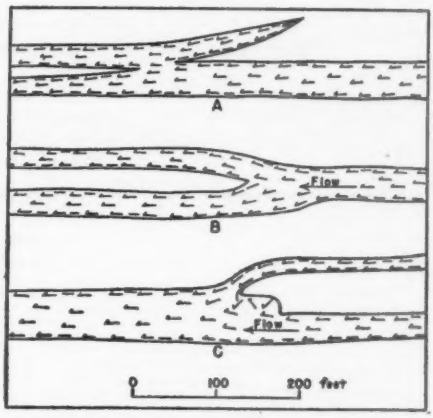

FIGURE 3.—SPLITTING OR MERGING OF SILLS

Sketched from geologic map. A, merging of two sills along a line about parallel to the direction of flow; flow is away from observer. B, splitting in the direction of flow. C, merging in the direction of flow.

flat-topped domes (or ridges?) with relatively steep slopes (Fig. 2-C), but in the Eagle River porphyry sills they are less definite in pattern (Fig. 2-D). Sills of the Lincoln and Eagle River porphyries as a group range from 1 to 400 feet thick, and individual sills may range from 50 to 400 feet, but most of the sills average 100–150 feet thick.

The ground plan of many of the sills is indeterminate for lack of exposures in the third dimension. Those for which data are available appear to be decidedly irregular. Some Elk Mountain porphyry sills appear to be amoeboid in shape, with long tongues protruding from a main body that is irregularly lobate. Others are, in part at least, markedly angular, with long, straight boundaries intersecting at large and small angles, like part of a crudely drawn star. Some sills, probably including the main sill of Pandro porphyry, have one straight edge marked by a dike, but in all the examples in the Pandro area the opposite edge has been destroyed by erosion. Most of the sills appear to have a roughly circular, semi-circular,

elliptical, or kidney-shaped main body from which project tongues, arms, or lobes of various shapes and sizes.

the character or structure of the wallrocks, or to local irregularities or discordance of the sills. Many of the sills change position with respect

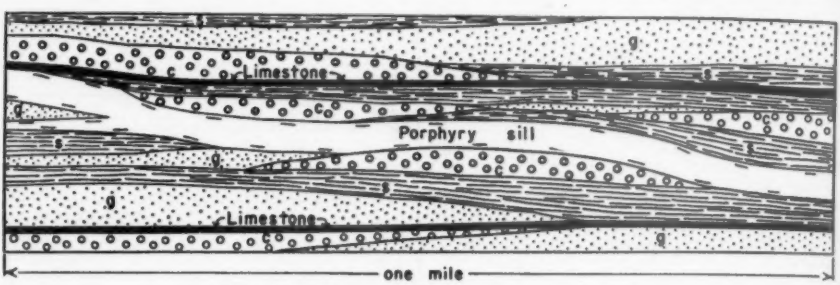

FIGURE 4.—STRATIGRAPHIC MIGRATION OF A SILL

Diagrammatic section shows migration with respect to persistent limestone beds in a section of lenticular clastic rocks. s, shale; g, grit; c, conglomerate.

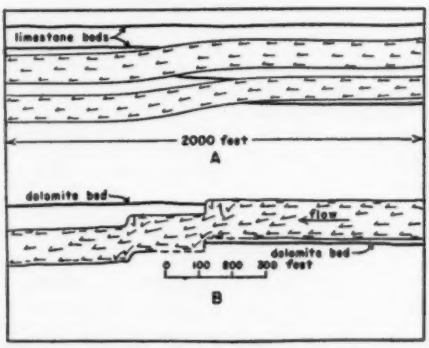

FIGURE 5.—MIGRATION OF SILLS BY ROLLING AND STEPPING

Sketched from geologic map. A, discordant roll in sills of Elk Mountain porphyry on west slope of Elk Mountain; flow at high angle to section, toward observer. B, steps in sill of Lincoln porphyry on north side of upper Pearl Creek; arrow shows component of flow in plane of section.

Many sills split or merge. In some, the line of "splitting" is about parallel to the line of flow and marks the merger of two sills intruded at slightly different horizons (Fig. 3-A). In others, the line of "splitting" is at high angles to the direction of flow, and one sill may split into two (Fig. 3-B), or, rarely, two sills may merge into one (Fig. 3-C) in going in this direction.

Changes in Position

The sills of the Pando area change stratigraphic position in several ways due either to

to persistent beds such as limestones because numerous short lenticular beds are present in the sedimentary section. Upon reaching the edge of a massive lens, an advancing sill might go either above or below the lens, or split around it, thus migrating stratigraphically without actually transecting the bedding (Fig. 4). Several sills change position at pre-porphyry faults. Some extend uninterruptedly across faults with no change in form; some change in thickness at faults; and some are deflected at faults but go to new stratigraphic positions, with or without a change in thickness.

Some sills change horizon by cutting across the bedding at a very low angle. Discordance of this type is generally not noticeable in small areas unless exposures are exceptionally good, or marker beds are present near the sills. At several places, sills gradually converge with, and then cross, continuous beds of limestone, thus showing a small discordance or low-angle "roll" for distances of a few hundred feet (Fig. 5-A). Some sills change position abruptly by means of steps or offsets which are essentially vertical. The line of offset may be at high angles to the line of flow, or it may be about parallel. In Figure 5-B, the line of offset is approximately at right angles to the plane of section, and the direction of flow is at an angle of about 30° to the section. A dike-like connection between the offset horizons presumably existed momentarily during intrusion. Such dikes might occupy either pre-existent frac-

tures or breaks related to the sill front, but after spread of the sill had started at the new horizon, identity of the dike would be lost. Offsets along a line parallel to the direction of stoping. At Green Mountain, 30 miles north of the Pando area, several hundred feet of sedimentary rocks are missing from between the thickest parts of superimposed domed sills,

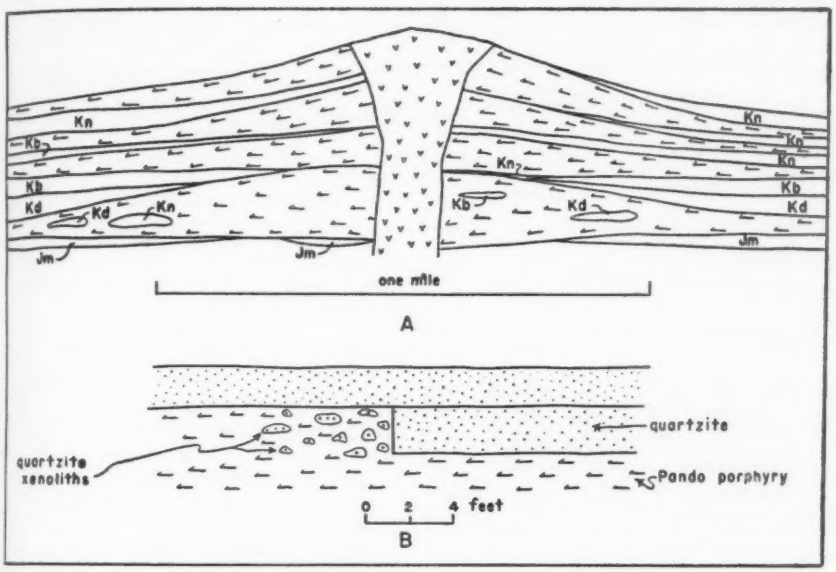

FIGURE 6.—DISCORDANT RELATIONS OF SILLS

A, diagrammatic section of Green Mountain, showing thinning of sedimentary formations between domed sills; Jm, Morrison formation, Kd, Dakota sandstone, Kb, Benton shale, Kn, Niobrara limestone. B, xenoliths of quartzite at end of truncated quartzite bed.

the flow are shown in Figure 8. Similar but smaller-scale offsets have been described from the Highwood Mountains, Montana, by Griggs (Hurlbut and Griggs, 1939, p. 1102–1104), who shows by sketches how the offsets might result from lateral coalesence of narrow fingers that advanced at slightly different horizons at the sill front. The offsets observed in the Pando area rarely affect the structure of the wallrocks; the sedimentary beds end abruptly at the vertical contact without distortion, and, although there may be a fracture in the wallrocks along the extension of the offset, there is little if any displacement on it.

At places sills have removed some of the adjoining host rocks and change horizon accordingly. Where thick, and particularly domed, sills are closely spaced, the sedimentary section is thinned markedly, probably both by squeezing out of shales by plastic flow and by removal of less plastic rocks by magmatic although present a short distance away where the sills are thinner, and large blocks of some of the missing rocks occur as xenoliths in the sills (Fig. 6-A). Apparently the Benton shale, normally about 350 feet thick, and the shaly parts of the Dakota sandstone and Niobrara limestone were squeezed out from between the domed sills, and at least a part of the sandstone and limestone was removed by stoping. Similar processes evidently operated at places in the Pando area, although on a smaller scale. In the area on the north side of the East Fork of Eagle River where the sills are closely spaced and thickest, the rock between the sills is almost entirely quartzite, although this lower part of the Pennsylvanian section normally contains a large amount of shale, suggesting that the shaly fraction was largely squeezed out from between the thickest parts of the slightly domed sills. Here as well as at a few other places in the Pando area, some beds of

quartzite and grit end abruptly against the sills, as if cut out by the porphyry. Unlike the steps or offsets, such broken beds do not reappear at the other contact of the sill. At one place, a swarm of quartzite xenoliths was observed opposite the end of a truncated quartzite bed (Fig. 6-B). Similarly, in Figure 2-D, the lenses of quartzite opposite the reentrant angle forming a split in the sill seem to be part of a once-continuous bed, and part of this bed has obviously been carried away by the porphyry magma.

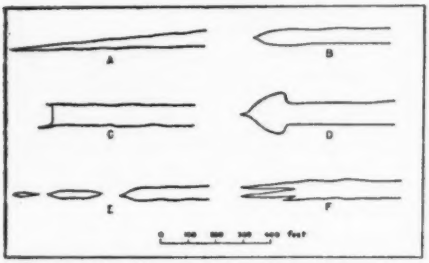

FIGURE 7.—PROFILES OF SILL TERMINATIONS

Termination

When the edge or "end" of a large sill is found, there is usually no way of ascertaining its relation to the gross shape of the sill, whether on the periphery of a circle, on the side or end of a tongue-shaped mass, or on the edge of some irregular mass. Consequently, in the discussion that follows, "termination" means the edge of the sill at any point on the periphery. The terminations take several forms in cross-section, apparently dependent in part on the character of the wall rocks and in part on properties of the individual porphyry magmas.

The long, slender wedge type of termination (Fig. 7-A) characterizes the edges of sills of Pando porphyry and is found also in some sills of Elk Mountain porphyry and some of trachyte porphyry at Green Mountain. The blunt lens type (Fig. 7-B) is fairly typical of sills of Elk Mountain porphyry and occurs in a few of Lincoln porphyry. A vertical edge (Fig. 7-C) was observed in one sill of Elk Mountain porphyry, with the main component of flow toward the vertical edge, not parallel to it. The bulging edge (Fig. 7-D) has been observed in a few sills of the Elk Mountain and Eagle River porphyries, and in some of these it is markedly asymmetric. The driblet and branching terminations (Figs. 7-E, 7-F)

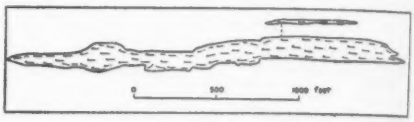

FIGURE 8.—IRREGULARITIES NEAR THE EDGE OF A SILL

Cross-section sketch of a tongue of a sill of Elk Mountain porphyry, on the east side of the divide, 2 miles north of Elk Mountain. Vertical scale slightly exaggerated.

are poorly represented in two or three sills of the Lincoln and Eagle River porphyries in the Pando area, but they are more common in other parts of central Colorado, as at Green Mountain and Leadville, and in the Beaver-Tarryall area, where Singewald (1942, p. 17–18) noted "intricate and ragged" terminations of monzonite porphyry sills.

Sills of all the porphyries, and especially those of Elk Mountain porphyry, are more irregular near the edges than in the central parts. Irregularities such as steps or offsets sudden bulging or pinching, discordant relations, off-shooting dikes, and wavy, warped contact surfaces may occur locally in any part of a sill, but near the edges of sills, and particularly near the ends of tongues of Elk Mountain porphyry, they occur together (Fig. 8), and their presence is a fairly reliable indication that the sill does not extend much farther. The concentration of such irregularities near the edges of sills or near the ends of tongues is probably due to the more rapid cooling and hence higher viscosity and early solidification of the border zone. At many places where the irregularities are prominent, the porphyry seems to have moved as a solid or plastic rather than liquid mass, as shown by contact features described below. Such a solid or highly viscous mass plugs the sill horizon and forces the fluid magma advancing from behind to seek new levels, or the almost solid mass may be literally driven into the rocks like

a wedge, crumpling slightly in the process and developing the warped surfaces characteristic of the edges of sills.

are typically more complex in the Pando and Elk Mountain porphyries than in the Lincoln and Eagle River porphyries. In the former

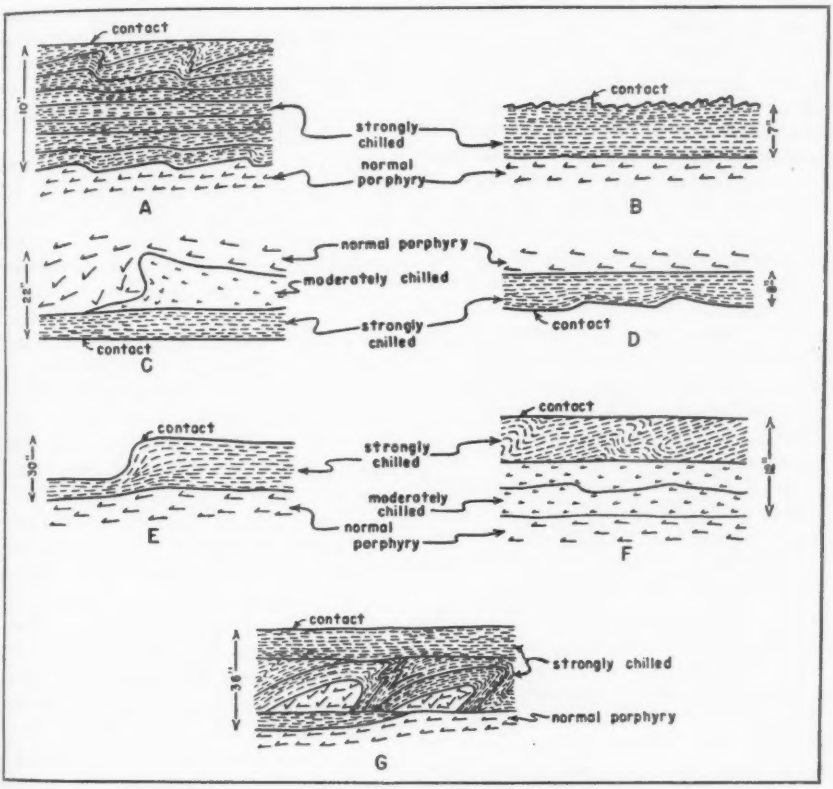

FIGURE 9.—TYPES OF DRAG FOLDS NEAR SILL CONTACTS

STRUCTURE

General

The sills show structural features of two general but closely related types, both produced by flow of magma. One type comprises planar and linear structures defined by the dimensional orientation of mineral grains. The other includes folds and fractures of various types visible without the aid of mineral orientation studies, here referred to as gross structures to distinguish them from those based on mineral orientation. The mineral orientation structures group, linear mineral orientation near the contacts is at right angles to that within the body of the sills; in the latter group, linear orientation is normally uniform from contact to contact, although the double-linear type of orientation may be present locally. Similarly, sills of the Pando and Elk Mountain porphyries show many types of folds and fractures of intrusive age, and those of the Lincoln and Eagle River porphyry show few, except for the one variety of Lincoln porphyry similar in occurrence to the Elk Mountain porphyry. The sills thus fall structurally into two groups.

Sills of the Pando and Elk Mountain Porphyries

Gross structures.—Sills of the Pando and Elk Mountain porphyries exhibit several kinds of structural features in addition to those based on mineral parallelism, such as drag folds in chilled or flow-banded porphyry, sharp or "faulted" contacts between chilled and unchilled porphyry, breccia, platy parting, zoning with respect to type and intensity of fracturing, stratification of phenocrysts, and various types of markings on the chilled contact surfaces.

Drag folds (Fig. 9) are conspicuous at places in the chill zones, where they are outlined by flow banding or by the shape of the chill zone as a whole, or by shape of either contact surface of the chill zone. Some of the folds take the form of crenulations in the contact surface, but many are independent of the contact and are truncated by it or distorted along it. At some places, minor folds in the porphyry at the contact surface are disconformable with larger folds deeper in the chill zone. At others, a layer of unfolded, banded and glassy-appearing porphyry an inch or two thick lies between the contact and the folded porphyry in the body of the chill zone, and the two layers are locally separated by a thin seam of pulverulent gouge. Rarely, spindle-shaped masses of tightly knotted ropy porphyry lie at the contact. These masses are embedded in the chilled porphyry but are not attached to it and are utterly independent of the adjacent structure. The spindles lie with their long axes parallel to the axis of drag folding and were clearly rolled along the contact for some distance. Except for small crenulations with an amplitude of less than an inch, drag folds in the contact surface itself are relatively rare, and the drag folds discussed refer to folds in porphyry and not in the contact surface unless specified. Drag folds in the chill zone range from almost microscopic crinkles to folds a few feet high, and from only slightly asymmetric to recumbent and thrust-faulted folds. The chill zones are relatively thin, mostly 6 to 12 inches thick, but where drag folding is strong the thickness varies directly with the size of the folds and generally with the intensity of folding. In some places, the chill zones are greatly thickened because of the piling up of pinched-off drag folds. Surfaces separating such folds or fold segments are faults of intrusive age and are themselves kinked or drag-folded in places.

Drag folds within the bodies of the sills are confined to rather narrow zones near the chill zones and are generally more open and of larger amplitude than the folds in the chill zones. They are defined by vague flow banding discernible at a few places or by mineral orientation, but they are weak and rare.

The chill zones are in general sharply separated from the unchilled porphyry. At many places the chilled and unchilled rocks are separated by a knife-edge contact or by a facture or parting surface. At most places where a gradation does occur, it takes place in a layer as little as half an inch thick or is distributed through several distinct layers each with uniform grain and separated from adjoining layers of different grain size by knife-edge contacts or parting surfaces. The abrupt change in texture at parting surfaces some of which are accompanied by a peculiar knobby or rounded breccia, shows differential movement between the chill zones and the body of the sill. The truncated drag folds at the inside contact of the chill zone and the lack of continuity between drag folds in the chill zone and in the adjacent part of the sill is other evidence. The parting surfaces or discontinuities at the inner edges of the chill zones are thus intrusive-stage flat faults, parallel to the contact and to the zoning within the sill and analogous to bedding faults.

Slightly younger intrusive-stage faults evidently formed after the outer part of the main

PLATE 1.—STRUCTURES IN SILLS

FIGURE 1. PLATY PARTING IN SILL OF PANDO PORPHYRY

Slightly brecciated area below hammer. Head of hammer at upper contact of sill.

FIGURE 2. STREAKING ON UPPER CONTACT SURFACE OF A SILL OF PANDO PORPHYRY

FIGURE 3. DRAG FOLD IN CHILLED PANDO PORPHYRY

Flow bands above pencil are bent sharply downward in fold to the right; lower part of fold is cut off by flat fault to right of pencil. Massive rock at upper right is quartzite.

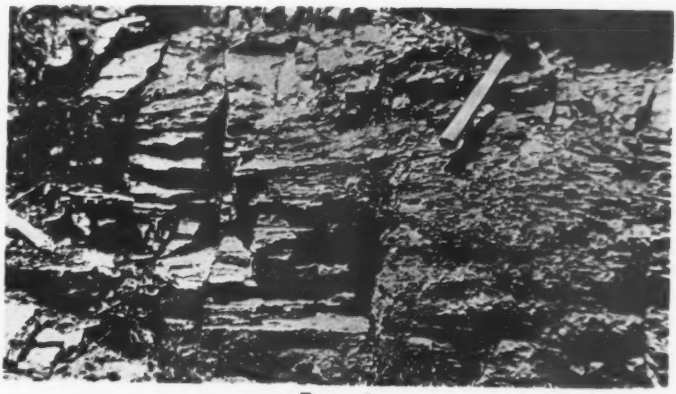

FIGURE 1

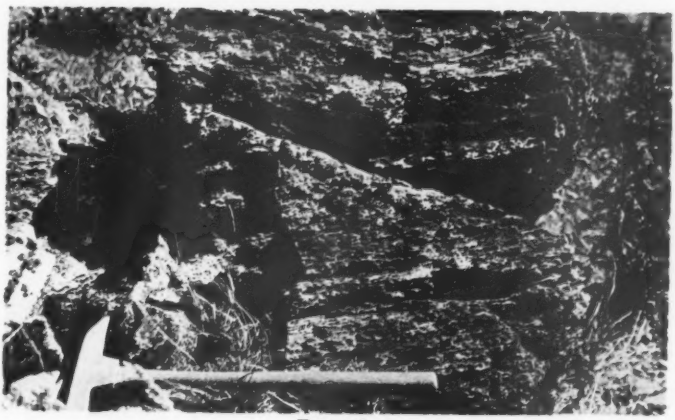

FIGURE 2

FIGURE 3

STRUCTURES IN SILLS

body of the sill had begun to solidify, for they are most prominent near the chill zones, where some are slickensided or coated with a thin sheath of gneissic material, and flatten or die out inward. As with similar fractures discussed by Balk (1937, p. 110–111, 117), the strike of the lineation on the fracture surfaces is parallel to the strike of the linear flow structure. Most of the fractures near the upper contact dip gently upstream with respect to the direction of magma flow, and those near the lower contact dip downstream. Assuming that the fractures result from continued forward movement of the magma, they are normal faults. A few small faults of about the same age but of opposite orientation, and thus presumably thrust faults, are present locally, particularly in and near contorted and lumpy chill zones, and older thrust faults related in origin to the drag folds are also found.

At some places, particularly in Pando sills the porphyry adjacent to the chill zones appears knobby and brecciated in zones up to 3 feet thick. Layers of similar breccia may recur farther within the sill but seldom more than 25 feet from the contact. The knobby breccia grades into a prominent platy parting that is parallel to the planar mineral orientation (Pl. 1, fig. 1). This parting has doubtless been accentuated by post-intrusion movements and is significantly prominent in segments of the Pando porphyry sill squeezed between thick sills of the younger Lincoln porphyry, but the parting surfaces are believed to mark planes of discontinuity resulting from differential movement between thin laminar sheets during intrusion. The sparse phenocrysts in the Pando porphyry are slightly concentrated between the parting surfaces and do not project across the partings. The outer parts of some sills of Elk Mountain porphyry appear stratified because the abundant rounded quartz phenocrysts are concentrated in well defined single-grain layers separated by prominent platy partings. Dozens of layers may be distinguishable through thicknesses of as much as 10 feet. The quartz grains appear to have been forced into layers by differential laminar flow, and the pronounced platy parting in the outer parts of the sills probably represents a similar but smaller scale movement.

Parts of some sills have a zoned appearance due primarily to the fracture pattern and density. Downward or upward from the generally massive central portion of the sill, a somewhat slabby zone with horizontal parting surfaces may be evident. This grades outward into a platy zone with more closely spaced partings and perhaps cut by the gently dipping intrusive-stage faults. Then may come a zone of rounded, knobby breccia, and then the darker and generally massive-looking, though possibly folded, chill zone.

The outer contact surfaces of the chill zones, and locally some of the platy parting surfaces near and parallel to them, characteristically show linear markings of some kind and at many places show two lineations at right angles. The lineations are of two orientations but three kinds and origins. One is a combination gneissic and slickensided structure oriented in the direction of magma flow and produced by friction between the sill and the host rock. The other two, oriented at right angles to the direction of advance, differ in their asymmetry with respect to the direction of magma flow. One comprises drag folds and features related to them; the other comprises furrows of opposite orientation from the drag folds, produced by intrusive stage faulting.

The drag-fold type of lineation on contact surfaces of Elk Mountain porphyry sills generally takes the form of small crenulations (Fig. 9-B) or more widely spaced troughs and ridges (Fig. 9-D). Contact surfaces of Pando porphyry sills are characterized by a streaking or grooving of somewhat different appearance (Pl. 1, fig. 2) though similar origin. The streaky surfaces somewhat resemble slickensided surfaces, but the lineation is of opposite orientation, parallel to the axis of drag folding and normal to the direction of sill spread. It is caused primarily by exposure of fine-scale flow banding in partly truncated gentle folds and to a lesser extent by small to minute drag folds. Rounded quartz phenocrysts affect the streaking in several different ways, but the pattern is usually consistent in any one locality. In some outcrops the phenocrysts merely interrupt the streaking without causing distortion. In some, the laminar streaking bows around the quartz grain on the "lee" side in terms of

direction of magma flow; in others it bends around the "stoss" side, or both sides, producing cross-wise eyes in the streaked surface. At some places the quartz grains have rolled for several inches, leaving a trail of bent and

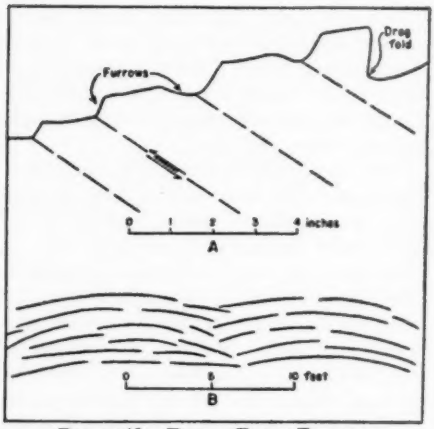

FIGURE 10.—THRUST-FAULT FURROWS AT SILL CONTACT

Near edge of sill of Elk Mountain porphyry. A, section; B, plan.

mangled laminae approximately at right angles to the direction of streaking. These features are obviously caused by differential movement between the quartz grain and the porphyry matrix, and the direction of bending in the laminar streaks depends on the relative movement between the quartz grain and the matrix.

The furrows with steep sides facing opposite to those of drag folds are prominent on contact surfaces near the edges of Elk Mountain sills. They are markedly asymmetric (Fig. 10-A) and take the form of a succession of small steps in the contact surface, with shallow troughs at the bases of many of the steps. The ridges between the furrows are typically coated with a thin shell of highly gneissic material and are deeply scored. Lineation of the gneissic shell and scored grooves is approximately at right angles to the general trend of the furrows, which form coalescing series of concentric arcs (Fig. 10-B). The furrows mark small thrust faults that displace the original contact surface, forming a series of steps, and they evidently developed in the hardening sill front during intrusion as a result of the push of more fluid magma behind. Each fault is marked by fine-grained, scaly breccia where it intersects the contact; differential weathering of this breccia and abrasion during the last movement of the

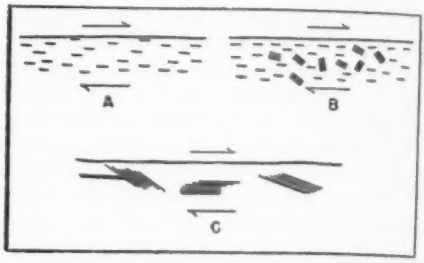

FIGURE 11.—ORIENTATIONS OF BIOTITE

As viewed in section, just below the upper chill zone. A, biotite flakes; B, biotite books and flakes; C, distorted biotite books.

sill probably account for the shallow troughs at the bases of the steps. The individual thrust faults can seldom be traced beyond the chill zones, but a weak planar mineral orientation noted locally in unchilled prophyry near the chill zones is about parallel to the thrust surfaces and may reflect them.

Many of the gross structural features suggest differential movement near the contacts of the solidifying sills. The body of a sill moved differentially with respect to the chill zones, as shown by the discontinuity at the inner edges of the chill zones, but the chill zones also moved, as shown by the drag folds in them and the discontinuity at the outer contacts. The inner parts of some sills moved differentially with respect to border zones a few feet thick immediately adjacent to the chill zones, and a knobby breccia developed in the plastic rock. Finally, differential movement evidently occurred between thin laminar sheets within all the differentially moving larger units, but this movement gradually died out inward.

Mineral orientation.—Within the sills, excluding contact zones from 1 to 25 feet thick depending on the individual sill, the planar mineral orientation is approximately parallel to the contacts, and the linear elements are oriented in the direction of magmatic movement, assuming the movement was per-

pendicular to the axes of drag folds at the contacts. Slight variations in the planar orientation in some sills suggest a wavy structure in the central part of the sill,—successive gentle anticlinal and synclinal flexures with

affect through cooling on viscosity, but it is expressed primarily by the differential movement between laminar plates within the sill, particularly near the contacts. Any laminar plate in a series of differentially moving plates

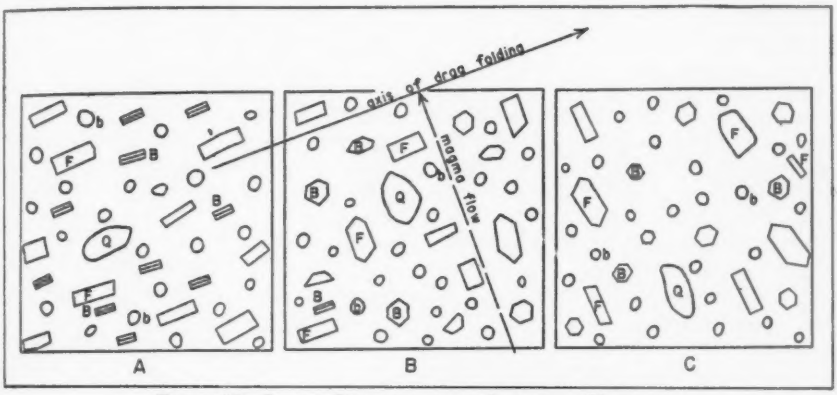

FIGURE 12.—LINEAR ORIENTATIONS AT SUCCESSIVE HORIZONS

A, adjacent to inner contact of chill zone; B, 5 feet inward from chill zone; C, 12 feet from edge of chill zone. Frequency of inequant grains exaggerated. F, feldspar; B, biotite book; b, biotite flake; Q, quartz.

axes at high angles to the direction of flow.

Planar orientation at or near the contact is parallel to the contact or to drag folds in the contact zone. Linear orientation in the chill zone and in the unchilled porphyry immediately adjacent to it is parallel to the axes of drag folds and normal to the orientation in the inner part of the sill and to the direction of magmatic flow.

The zone of transition between what might be called the contact lineation and the internal lineation is 1–25 feet thick and is characterized by two linear mineral orientations at right angles. In a zone a few to several inches thick in the unchilled porphyry adjacent to the chill zone, the contact lineation predominates almost to the exclusion of the internal lineation. Going inward, the internal lineation becomes progressively more prominent as the contact lineation becomes weaker and finally disappears.

The co-existence of two linear structures suggest two independent orienting forces, but the primary orienting force is the flow of the magma. The other force is a resistance to flow traceable ultimately to the wall rocks and their

of viscous material is subjected to rotational stress. Two mineral orientations can exist in the stress field because mineral grains first oriented by simple flow respond differently to rotational stress depending on their shape. Biotite flakes tend to remain in the plane of the original planar flow structure and are the best means of distinguishing the planar orientation (Fig. 11-A). Biotite books and feldspar grains tend to rotate in the fields of shearing couples set up near the outer edges of the sills; this brings their edges into the plane of the true planar structure, defining a linear orientation at right angles to the direction of flow (Fig. 11-B). Many of the rotated grains dip subparallel to the plunge of the internal lineation, but enough are rotated to other positions to make the secondary planar structure vague or indiscernible. Occasional biotite books in the outer part of the structural transition zone are distorted by gliding on the cleavage (Fig. 11-C). In the chill zone and immediately inward, almost all the linear elements in the plane of the true planar structure are aligned parallel to the axis of drag folding (Fig. 12-A). A few

feet inward from the chill zone, some of the elongated grains are aligned normal to this axis and some are parallel to it (Fig. 12-B). A few feet deeper within the sill, the average orientation of the elongated grains is at right angles to the axis of drag folding; this is the internal lineation (Fig. 12-C). In the inner part of the structural transition zone, the two linear orientations tend to alternate. The contact lineation may be found only along certain planes of platy fracture which are presumably contact planes between zones that moved differentially. It may be conspicuous on such a surface and unrecognizable an inch above or below.

Rotation of hexagonal or subcircular biotite books so as to bring any edge into the plane of the true planar structure will produce a maximum or near-maximum linear exposure of the biotite. Simple rotation of tabular feldspar crystals in the field of a shearing couple acting parallel to the direction of elongation of the crystals should up-end them, producing an elongation in the plane of the true planar structure that is not a maximum for such crystals. Feldspar crystals near the contact are nevertheless oriented with maximum elongation in the direction of the contact lineation. Simple endwise rotation is apparently accompanied by a sideward roll to a more stable position, a tendency promoted by the terminal asymmetry of the crystals.

The direction of spread of the sill magma is the principal direction of movement, to use Cloos's terminology (1946, p. 25, 45); the direction of drag-fold axes and of the linear mineral orientation near the contacts, a consequence of movement in the principal direction, is the subordinate direction. Thus lineation is parallel to the principal direction and to the *a* axis within the bodies of the sills and near the contacts is parallel to the subordinate direction and the *b* axis, opposite that found in most igneous bodies, including many sills. Cloos (1946, p. 25) says "Generally, it seems as though intense friction along walls or in small plutons tends to orient flow lines into the direction of motion". In sills of the Elk Mountain and Pando porphyries, however, a layering produced by differential movement between laiminar plates controlled development of structure near the contacts. Thus the outer parts of the sills behaved like bedded rocks; in response to a force in *a*, they developed folds with axes in *b*, and linear parallelism of mineral grains developed as a result of rotation about this axis.

Examples of structural features.—A sill of Elk Mountain porphyry 1400 feet north-northeast of Elk Mountain affords a good example of these structures. The chilled upper contact surface has a streaked or grooved appearance caused by tiny crenulations and small drag folds from a fourth to one inch high in the chilled porphyry. The axes of the folds plunge N. 30° E., 19°. The southeast limbs of the drag folds are vertical or overturned. In unchilled porphyry immediately under the chill zone, the planar structure strikes N. 10° W. and dips 29° E.; the linear mineral structure, like the drag folds and crenulations of the contact, plunges N. 30° E., 19°, but there is a faint suggestion of a linear orientation plunging S. 60°E. Drag folds are absent in this zone. Downward into the sill, the S. 60° E. orientation becomes stronger as its complement becomes weaker, and at about 15 feet below the contact the only linear orientation present plunges S. 60° E., 23°. The planar orientation remains constant, striking N. 10° W., and dipping 29° E. These orientations obtain throughout the remaining 60 feet of the sill exposed at this locality. Several hundred feet to the southwest, where the lower 50 feet of the sill is partly exposed, the planar structure 20 feet above the contact strikes N. 10° W. and dips 26° E., and the linear structure plunges S. 70° E., 23°. Three feet above the contact, the predominating linear orientation plunges N. 20° E.

Mineral orientation is not so easily determined in the Pando as in the Elk Mountain porphyry, but drag folds, streaking on the contact surface, and evidence of differential laminar movement are better shown. About 3 miles southeast of Pando and 2800 feet east of the ranch house near the forks of Eagle River, the upper contact of the main sill of Pando porphyry is well exposed beneath a small cliff of massive, white Pennsylvanian quartzite. The contact is wavy but on the average strikes north and dips 6°–10° E. Two inches of glassy-

looking, finely banded rock at the contact is followed successively by 6–12 inches of knobby, dense chilled porphyry, about 8 inches of banded fine-grained chilled porphyry, and a 6- to 10-inch band of quartz-phenocryst-rich chilled porphyry showing drag folds and faults (Pl. 1, fig. 3). Normal Pando porphyry that is only slightly chilled lies below the drag-folded zone. The axis of the drag fold illustrated plunges N. 50° E., 8°. The linear mineral orientation on the fold and in undeformed porphyry for a few feet below the fold also plunges N. 50° E., 8°. About 10 feet below the fold, the linear structure plunges S. 40° E. at a low angle. The drag fold is cut off at the base by a flat intrusive-stage fault; between this fault and a second fault that dips gently southward a few inches below it is a flat-lying streak of the drag-folded porphyry which was doubled back by folding and carried northwestward by faulting. Both fault planes are slightly kinked parallel to the axis of the drag fold. The porphyry below the zone of drag folding, and only a short distance below the contact in other parts of the outcrop where the chill zone is thin, has a prominent platy parting (Pl. 1, Fig. 1), which in many places grades into a knobby breccia. The top surface of the sill is poorly exposed at this locality, but in other outcrops near by shows prominent streaking. The streaking shown in Figure 2 of Plate 1 plunges N. 75° E., the direction of the linear mineral orientation near the contact. About 8 feet below the contact, the linear orientation plunges S. 15° E.

Sills of the Lincoln and Eagle River Porphyries

Gross structures.—Except in one variety of Lincoln porphyry that resembles the Elk Mountain porphyry in habit, gross structural features such as those found in the sills of the Pando and Elk Mountain porphyries are absent or present only locally in sills of the Lincoln and Eagle River porphyries. Drag folds have not been observed in the Eagle River porphyry, and most of the few folds found in Lincoln porphyry sills are reflected in the wallrocks, in contrast to the folding independent of the contacts in sills of the Pando and Elk Mountain porphyries. Such folds are generally apparent only in the shape of the contact, and a corresponding folded structure in the adjacent porphyry is rarely discernible. Most of the larger folds seem to have resulted from relatively large-scale doming of the sills rather than from the local drag effect of the walls. Drag folds independent of the contact are found only rarely. Most of them are broad and relatively symmetrical and are weakly defined.

Sharp contact between chilled and unchilled porphyry, such as occurs in sills of the Pando and Elk Mountain porphyries, is absent in Eagle River porphyry sills and has been found at few places in sills of Lincoln porphyry. Chilled Lincoln porphyry typically grades gradually into unchilled porphyry. Similarly, breccia of the type found at places in many sills of the Pando and Elk Mountain porphyries has not been observed in Eagle River porphyry sills and is very rare in Lincoln porphyry sills. Stratification and platy parting have not been observed in either Lincoln or Eagle River porphyry sills. Grooves of the slickenside type, at places accompanied by gneissic structure caused by granulation and drawing out of mineral grains, are present locally on contact surfaces of Lincoln porphyry, but lineations such as drag folds and thrust-fault furrows at right angles to the direction of flow are extremely rare. The grooves found on unchilled as well as chilled contact surfaces evidently resulted from scoring action of the wallrocks during movement late in the period of consolidation.

Mineral orientation.—Planar and linear mineral orientations in general are uniform from contact to contact in sills of the Lincoln and Eagle River porphyries. They are poorly defined in the central part of the sills, but the average orientations of biotite flakes and the larger phenocrysts give more or less consistent results in a given locality. They are better defined near the contacts. The planar orientation is about parallel to the plane of the contact except in a few places where large drag folds in the contact are reflected to a depth of a few feet in the sill. Linear orientation is parallel to the direction of flow as determined from the few drag folds and the relation to the intrusive center, except locally near the contacts of some Lincoln sills where orientation at right

angles to the direction of flow is found. In contrast to sills of the Pando and Elk Mountain porphyries, however, such an orientation is found in zones only a few inches thick adjacent to the contacts and is spotty in occurrence.

TABLE 1.—STRUCTURAL CHARACTERISTICS OF THE TWO GROUPS OF SILLS

Sills of Pando and Elk Mountain porphyries	Sills of Lincoln and Eagle River porphyries
1. Linear mineral orientation in zones up to 25 feet thick near contacts at right angles to orientation in central part of sill, and to direction of magma flow.	1. Linear mineral orientation uniform across any section of a sill, and parallel to direction of flow, except locally in thin zones near contacts of some Lincoln porphyry sills where orientation is at right angles to flow.
2. Drag folds in porphyry, particularly in chill zones; generally not reflected in contact.	2. Drag folds absent in Eagle River porphyry and uncommon in Lincoln porphyry; generally reflected in contact.
3. Sharp or "faulted" contacts between chilled and unchilled porphyry, except locally in some sills.	3. Gradational contacts between chilled and unchilled porphyry, except locally in some sills.
4. Knobby breccia in zones near and parallel to contacts in some sills.	4. Breccia absent in Eagle River porphyry and rare in Lincoln porphyry.
5. Parting surfaces of intrusive age cause conspicuous platy parting parallel to contacts, particularly in Pando porphyry.	5. Absent.
6. Local stratification near contacts, caused by concentration of phenocrysts in layers between platy partings.	6. Absent.
7. Sills locally zoned with respect to type and intensity of fracturing.	7. Absent.
8. On contact surfaces, combination gneissic and slickenside linear structure parallel to direction of flow, particularly in sills of Elk Mountain porphyry.	8. Present locally in sills of Lincoln porphyry.
9. On contact surfaces, lineation at right angles to direction of flow, produced by crinkles and small drag folds, especially in Pando porphyry.	9. Rare in Lincoln porphyry; absent in Eagle River porphyry.
10. On contact surfaces, furrows at right angles to direction of flow but oriented opposite to drag folds; produced by intrusive-stage thrust faulting.	10. Absent.

Difference in Structure

The pronounced difference in structure of the two groups of sills (Table 1) could reflect some difference in the magmas, the environment of intrusion, the intrusive history, or combination of these, but, actually, most of these differences appear to be a direct reflection of difference in the viscosity of the magmas. All the sills were intruded in the same area and into the same rocks at almost the same time. The intrusive histories appear to have been about the same. Distance traveled by the sill magma was not a controlling factor, for the structural habit of the sills is not observed to change with distance from the intrusive centers. Rate of flow might affect both form and structure, but, for sill magmas intruded in the same environment at almost the same time, rate of flow is controlled primarily by viscosity and the pressure gradient. Unless there was a great difference in the propelling forces for the two groups of porphyry magmas—a possibility for which no other evidence has been recognized—rate of flow should be a direct expression of viscosity.

Most of the structural features that dis-

tinguish sills of the Pando and Elk Mountain porphyries from those of the Lincoln and Eagle River porphyries imply a high viscosity. The folding and thrust faulting of the porphyry near the contacts during intrusion, and the rotation or rolling of mineral grains at depths of several feet from the contacts, could take place only in a highly viscous medium. The general absence of such features in sills of the Lincoln and Eagle River porphyries suggests that these magmas were inherently less viscous.

There is considerable supporting evidence of a difference in viscosity between the two groups of porphyry magmas. Quartz and feldspar phenocrysts in chilled Pando and Elk Mountain porphyries are more abundant and more nearly normal in size than those in the Lincoln and Eagle River porphyries, and the quartz grains show a greater degree of resorption, suggesting that the Pando and Elk Mountain porphyries were crystallized in greater degree at the time of intrusion than the Lincoln and Eagle River magmas, and thus where more viscous.

A difference in metamorphic effect on the wallrocks also suggests that the Pando and Elk Mountain magmas were coolor or dryer, or both, and thus more viscous. The Pando and Elk Mountain porphyry sills had no noticeable effect on quartzite and grit, caused only slight recrystallization and bleaching of limestone, hardened or baked black shale for a few inches, bleached and hardened red, slightly limy, arkosic shale for only 1 to 2 inches, and had no effect on micaceous shale. Metamorphosed zones range up to 4–5 feet thick adjacent to sills of Lincoln porphyry and up to 10 feet adjacent to those of Eagle River porphyry. Grit is altered to vitreous quartzite or arkosite; shales are altered to hornfels or to a highly sericitic rock; and limy shales are epidotized.

The weakness of chill zones in sills of Eagle River porphyry and the local absence of chill zones in sills of both the Lincoln and Eagle River porphyries indicate either that the magmas were intruded into warm rocks or that formation of chill zones was prevented by convection (Grout, 1932, p. 233). However, the Lincoln porphyry locally contains relatively wide chill zones, indicating that the wall rocks were not warm and thus that convection was the controlling factor in the development of chill zones, at least in the Lincoln porphyry. Convection suggests low viscosity, particularly as it could be effective for only a short time in small intrusives such as the sills, which, unless highly superheated, would probably lose enough heat to solidify within a year or two (Lovering, 1935, figs. 1, 2, 3).

A sub-pegmatite facies in a stock of Lincoln porphyry near Alma (Singewald, 1932, p. 63) and miarolitic cavities and vuggy epidotized inclusions in the Eagle River porphyry indicate relatively high content of volatiles and consequent low viscosity of these magmas.

The effective viscosity during emplacement is the sum of two factors: (1) the initial viscosity, dependent upon temperature, chemical and physical composition, and pressure, and (2) a viscosity factor caused by cooling at the contacts. The initial viscosity would be more or less constant for a given porphyry magma within the zone in which sills were emplaced. The factor caused by cooling might range widely, depending upon accidents of environment, speed of intrusion, and distance traveled. If the initial viscosity were high, the effective viscosity might be high enough to cause sills of a given porphyry consistently to develop structures characteristic of high viscosity. If the initial viscosity were low, the factor caused by cooling might be insufficient to raise the effective viscosity to the critical value above which structure resulting from high viscosity develops, or widely ranging values of the factor caused by cooling might cause this structure to be present in one place and absent in another. The Pando and Elk Mountain porphyries appear to have had relatively high initial viscosity, and they consistently developed complex sill structure. The Lincoln and Eagle River porphyries appear to have had relatively low initial viscosity, and they developed simple structure except locally where high values of the cooling factor caused complex structure to develop.

Mechanics of Intrusion

Features of Sill Spread

Pattern of flow.—In tongues at the margins of sills, particularly in those of Elk Mountain

porphyry, magma flow was not everywhere parallel to the axis of the tongue, like water in a pipe, but was in general outward toward the sides of the tongue (Fig. 13-A). This, amply demonstrated by the orientations of linear minerals, drag folds, thrust-fault furrows, and gneissic structure and slickensides on the contact surfaces, suggests that intrusion in the tongues begins as a rod-shaped protuberance of magma which later grows to a flat, tongue-shaped mass by lateral spread. The presence of tongues at the margins of sills, and the fact that they grow by lateral spread, suggests that the main bodies of the sills grow by coalescence of the advancing tongues at the margins, rather than spreading outward radially with an even front. Local variations in lineations near the edges of sills, particularly along the lines of projection of the tongues back into the body of the sill (Fig. 13-B) is supporting evidence of this process.

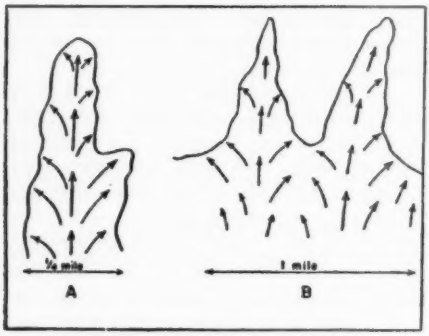

FIGURE 13.—PATTERN OF FLOW

A, in tongue of sill; B, in body of sill, at base of coalescing tongues.

Correlation between direction of plunge of a given lineation and direction of flow is low in gently dipping sills (Fig. 14). For many intrusive bodies, it is more or less justifiably assumed that the movement was from the direction of plunge, i.e., that movement of the magma had an upward component. Later tilting might affect the plunge, and the possibility of such tilting is generally taken into account. But many sills are gently inclined or almost horizontal. It is very difficult to determine whether they were intruded into gently tilted rocks or the entire section was tilted after intrusion. The plunge of the linear mineral orientation therefore may be meaningless as an indicator of the direction of flow, and the direction must be determined from other features such as drag folds. In some of the few sills in which primary plunge seems definitely to be a component of the observed plunge, there is evidence that the primary plunge was away from the source, or movement was in the down-plunge direction. This is true especially of the upper parts of thick doomed sills and of the rapidly pinching edges of sills. In the special case of superimposed domed sills, as at Green Mountain (Fig. 6-A), flow was obviously in the down-plunge direction.

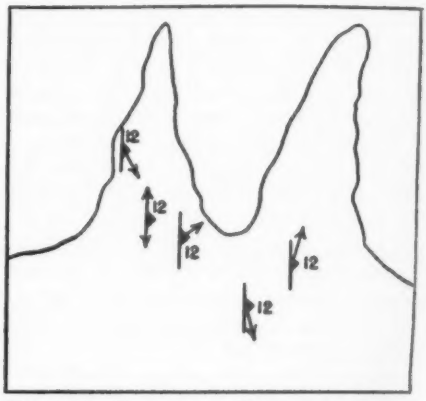

FIGURE 14.—VARIATIONS IN DIRECTION OF PLUNGE

Based on some of arrows in Figure 13-B. Arrows on the dip symbols show different directions of plunge of a given linear element after sill is gently tilted.

Character of flow.—Many of the sills show no special features, other than size, that give a clue to the character of the flow, whether it was slow or fast, steady or in spurts. The small size of some of the sills suggests that intrusion was relatively rapid—probably a matter of days rather than months or years—since, assuming that intrusion took place within 100° to 200° of the crystallization temperature and that the wallrocks were cool, as suggested by the presence of chill zones, the thin minor sills would lose enough heat to solidify in a few days, and an average sill about 20 meters thick would

probably solidify in a year (Lovering, 1935, figs. 1, 2, 3).

Spurts in the flow of magma are recorded by sets of chill zones within unchilled porphyry in some sills, particularly near the edges. Spurting is also suggested by small pods of porphyry apparently isolated a short distance beyond the sill front (Fig. 7-E) and perhaps by the thrust-fault furrows near the edges of sills.

Sills of the Pando and Elk Mountain porphyries, and somewhat less certainly those of Lincoln porphyry, apparently spread by quiet flow along a selected bedding plane. There seems to have been no advance invasion of the sill horizon by vapors or liquids, for there is no sign of disturbance or alteration at the horizon of the sill a few feet beyond the edge of the sill. The forward edge of the sill is typically wrapped in a casing of tough, kneaded clay or shale and bleached and altered. The intrusion breccia suggests that advance of the sill proper was preceded by explosive introduction of fluids or tenuous magma, and therefore that emplacement of the sill magma itself may have been very rapid. A small sill of quartz latite porphyry

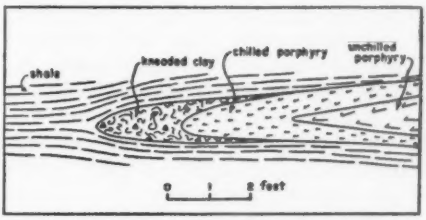

FIGURE 15.—CROSS SECTION OF THE EDGE OF A SILL

Small sill of Pando porphyry on cliffs northeast of loop in old highway, in valley of east Fork of Eagle River.

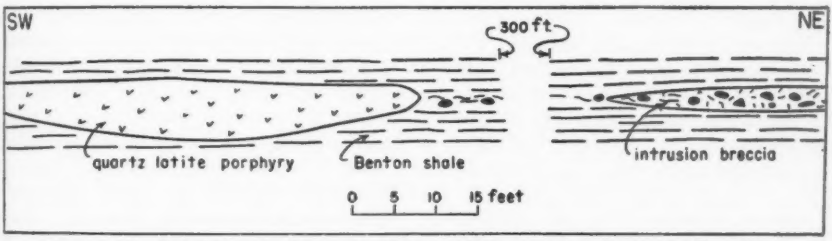

FIGURE 16.—CROSS-SECTION SKETCH OF SILLS OF PORPHYRY AND INTRUSION BRECCIA

Showing cross-cutting relations of edge of sill. Green Mountain area.

embedded rock fragments (Fig. 15), and this together with the chilled edge of the sill forms a blunt and solid wedge that was pushed along the bedding plane by the pressure of the liquid magma behind.

Sills of the Eagle River porphyry are in places accompanied by intrusion breccia which, although poorly exposed, seems to be an advance guard of the main sill. Similar breccia, in thin, irregular sills and branching dikes, has been observed beyond the edges of sills in the mines at Leadville and Breckenridge and in the Green Mountain area. The breccia may consist of fragments of country rock, many of them silicified, and chilled porphyry in a shaly matrix, or of rock fragments in a matrix of dirty, contaminated igneous material. At places the wallrocks of the breccia seams are and intrusion breccia that crosses the highway on the south side of Green Mountain reservoir 4000 feet east of the dam apparently was literally blown into place. It is 6 to 8 feet thick, and the bedding planes at the top and bottom of the sill maintain this same separation beyond the ends of the sill (Fig. 16), as if emplacement of the sill had entailed removal of an equal volume of Benton shale. The shale is baked for 10 feet below the sill and 15 feet above it. There are a few crenulations in the shale at the contact, but the shale is little disturbed. The porphyry is tightly frozen to the shale. The sill wedges out suddenly to the northeast and the tip is encased with chilled and contaminated black porphyry that contains scattered grains of quartz and plagioclase and grades into the light-colored and coarser-

grained quartz latite. A narrow band in the baked shale opposite the tip of the sill is slightly rusty and contains a few small fragments of the dark, chilled porphyry. About 300 feet northeast of the tip of the sill, the rusty zone widens into a lens of breccia about 3 feet thick. At some places the breccia consists of cracked lenses of the black, chilled porphyry in brecciated, baked shale, and in others it consists of fragments of baked shale, well-rounded boulders of chilled porphyry, and pebbles and boulders of strongly altered Precambrian rocks in a soft, iron-stained, clayey matrix that looks like altered latite. Similar features are found southwest of the sill, although the end of the sill itself is covered; lenticular seams and very irregular branching dikes of breccia and pods of fairly pure quartz latite porphyry occur in a streak of baked and rusty shale at the horizon of the sill, exposed in a road cut. Structure in the sill could not be determined with certainty, but there may be a very weak linear structure plunging 45° SE, about perpendicular to the exposure. The breccia, the presence of Precambrian rocks at this horizon several hundred feet above the basement rocks, and the cross-cutting relations of the ends of the sill all suggest that this sill was emplaced explosively.

Sills in Light of Popular Theory

When the two groups of sills of the Pando area are considered in light of the generally accepted theories of sills and laccoliths, a paradox at once becomes evident. According to popular theory (MacCarthy, 1925), relatively fluid magma forms sills, and viscous magma forms laccoliths; in sills themselves, the lower the viscosity, the greater the spread (Pirrson, 1898, p. 585). Uniform sills are supposed to be formed by magma of low viscosity, and the irregularities in many sills are popularly ascribed to somewhat higher viscosity (Emmons, Irving, and Loughlin, 1927, p. 53). In the Pando area, the sills of Pando and Elk Mountain porphyry are by far the most uniform and widespread, but they also show evidence of high viscosity. The sills of Lincoln and Eagle River porphyries appear to have been more fluid but are less widespread and more irregular, and some of the Lincoln porphyry sills grade into thick domed sills or laccolithic bodies.

The irregularities of the Lincoln and Eagle River porphyry sills might be accounted for by other features than relative viscosity, such as speed of intrusion, lenticularity of the host rocks, or sudden change in magmatic pressure during intrusion, but if these sill magmas of a given viscosity spread a given distance, the far wider spread of apparently much more viscous magma in uniform sills of Pando and Elk Mountain porphyry seems contrary to natural law. As Daly (1933, p. 77) has said, "Most writers assume an active role for the sill magma, the roof being lifted from the floor by liquid pressure—the laccolithic mechanism". If this is true, a sill of viscous magma should not extend itself as far as one of fluid magma, for (Paige, 1913, p. 541–549) high viscosity interferes with the hydrostatic law; with increasing viscosity due to cooling at the margin of the intrusive body, a point is finally reached when the magma cannot exert enough pressure at the margin to lift the load and extend the sill. Magma added after this stage must add thickness rather than area and thus tends to produce a domed sill or a laccolith. Moreover, the sill form greatly increases the area of contact of a body of magma, and, as contact friction increases with viscosity, the spreading of highly viscous magma as sills that must lift the overlying load should be greatly inhibited by contact friction as well as by loss of pressure. The pressure of a shallow body of relatively viscous magma should thus be relieved more easily by continued rise as a unit toward the surface than by spreading out in numerous sills.

Why, then, should the viscous Pando and Elk Mountain magmas form more widespread and uniform sills than the more fluid Lincoln and Eagle River magmas? Perhaps the answer lies in the stress environment during intrusion. The sills are in a comparatively narrow belt of sedimentary rocks between the great Sawatch and Park-Front Range uplifts of central Colorado. According to Lovering and Goddard (1938, fig. 2), they were intruded early in the period of northwest folding that was the first major event of the Laramide orogeny, and

before the succeeding period of northwest faulting (including thrusting). Crustal stresses therefore could have been important in their emplacement. The lifting action of beds stressed by lateral compression would reduce the load in certain, generally incompetent, stratigraphic zones, allowing sills to invade these zones without lifting the entire weight of the overlying rocks. Cross (1894, p. 236) suggested this for some of these same sills in the Kokomo or Tenmile district. Although made up of lenticular beds, the Pennsylvanian and Permian (?) host rocks of the sills contain several thick and persistent zones of grit and conglomerate that might transmit lateral forces over wide areas. Other competent beds, such as the Dakota sandstone, existed in the Mesozoic rocks that presumably overlay the area at the time of intrusion.

Considering the lenticularity of the host rocks, the uniformity of the Pando and Elk Mountain porphyry sills also suggests that intrusion was passive or "permissive". If a body of magma has the power to lift the overlying rocks to make room for a sill, the pressure exerted by the magma is necessarily greater than the static load pressure on the sedimentary rocks at the horizon of the sill. The sedimentary rocks, however, are in equilibrium with the load, unless some residual compaction is retained from an earlier period of greater burial or of deformation. In the Pando area, intrusion took place early in the Laramide revolution (Lovering and Goddard, 1938, fig. 2), before the rock had been deformed and only shortly after the close of Cretaceous sedimentation, when the load was at a maximum. The degree of compaction at a given horizon was thus determined by the static load, and the pressure greater than this load exerted by the wedge-like action of a sill forced into the rocks should cause increased compression of compactible beds such as shales. Thus in attempting to lift the load, the sill gains "space" by inducing a compactive yielding in the host rocks. Such yielding is well indicated where sills locally bulge downward (Figs. 2-D, 8), as if, in exerting an upward push, the base or foundation had yielded at weak spots, like mud beneath an auto jack. Most of the yielding naturally is upward, however, and "space" made by increased compaction of 2 or 3 miles of beds between a sill and the surface might allow emplacement of, say, a 100-foot sill with little or no compensating change in the position, or altitude, of the surface.

Compaction yielding of this type should be greatest in the first compactible (shaly) zone above the sill and should decrease rapidly upward, because any "space" gained by compaction in one shaly zone reduces the pressure transmitted to the base of the next overlying shaly zone. If the sill were overlain by persistent and even beds of alternating shales and sandstones, the amount of yielding would be nearly uniform over the area of the sill, assuming that pressure in the sill magma is uniform. But if the section above the sill is made up of lenticular beds of shale and sandstone or grit and therefore consists of widely ranging proportions of relatively compactible and incompactible rocks, as it does near Pando, the yielding would be nonuniform, and the sill would be nonuniform in thickness, pinching beneath parts of the section where little space could be gained by compaction and bulging below areas where large space could be gained.

Thus the irregularities such as pinching and swelling and small-scale doming characteristic of the sills of the Lincoln and Eagle River porphyries would seem normal for sills intruded into lenticular host rocks, and they suggest that the sills of these porphyries made room for themselves by lifting the overlying rocks. The weakness or absence of such irregularities in sills of Pando and Elk Mountain porphyries seem to support the conclusion that the sills were introduced passively, aided by the arching action of laterally compressed host rocks, and without the power to exploit to any great degree the weak spots in the nonuniform host rocks.

Although differential compaction of the host rocks may account for small-scale domes such as those on some sills of Lincoln porphyry, it probably would not account for large-scale doming or gradation into laccolithic bodies. In both the Pando area and at Green Mountain, the swellings to laccolithic proportions are at or near probable supply vents, and the laccolithic doming above these vents probably resulted from rapid supply of a large volume of magma. If a large volume of magma is supplied

rapidly through a vent, the rate of supply might easily exceed the rate of distribution in sills, where movement is necessarily slowed by friction along the greatly increased area of contact (Pirrson, 1898, p. 584–587; Paige, 1913, p. 545–549; MacCarthy, 1925, p. 7). Once the overlying rocks were domed they would protect the domed magma body from obliteration even if the supply ceased while the magma dome and sills remained liquid.

Conclusions

Many of the features of form and structure described result from special qualities of the magmas and the host rocks and may not be typical of sills in general; but, as the sill is one of the most neglected forms of igneous body with respect to structural investigation, there are few standards for comparison. More mafic sills might differ considerably in structure from those of quartz monzonite porphyry in the Pando area because of their inherently lower viscosity, and more siliceous sills, if they exist, might be much more complex. Because of the great difference in environment of intrusion, granitic and other coarse-grained plutonic sills in metamorphic host rocks may be expected to differ in many respects from the hypabyssal intrusives here considered. Sills in uniform host rocks such as the Cretaceous shales at Green Mountain differ in form from those in non-uniform rocks such as in the Pando area, and sills in undeformed rocks such as in the Colorado Plateau, to judge from the literature, differ in form and probably in structure from those intruded during mountain building, as in the Pando area.

In the Pando area, sills of four kinds of porphyries fall naturally into two structural groups which reflect differences in viscosities of the magmas. Sills of the Pando and Elk Mountain porphyries show intrusive-stage folding and faulting and, near the contacts, a linear mineral orientation at right angles to that in the inner parts of the sills and to the direction of magma flow. This lineation was produced by the rolling of mineral grains between closely spaced planes of differential movement in the outer parts of the sills. These structures suggest that the Pando and Elk Mountain porphyry magmas were relatively viscous. Sills of the Lincoln and Eagle River porphyries show little evidence of intrusive-stage folding and faulting and the linear mineral orientation is normally uniform from contact to contact, suggesting that these magmas were relatively fluid.

The porphyries that show evidence of high viscosity formed the most uniform and widespread sills; the relatively fluid ones formed the most irregular and least extensive sills as well as domed sills that grade into laccolithic bodies. This is just the reverse of what should be expected, and it is suggested that spread of the viscous Pando and Elk Mountain magmas in relatively thin, widespread, and uniform sills was aided by the lifting action of laterally compressed competent beds. Local irregularities and small-scale doming of sills of the Lincoln and Eagle River porphyries is attributed to differential compaction of the lenticular host rocks, and the large-scale doming or swelling to laccolithic proportions is attributed to rapid supply of magma near source vents.

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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SOME IMPLICATIONS FOR GEOPHYSICS OF HIGH-PRESSURE PHENOMENA*

By P. W. BRIDGMAN

(Address before The Geological Society of America, Sixty-third Annual Meeting)

Measurements have been made with modified apparatus of the viscosity of a number of methyl siloxanes under pressure. The new apparatus makes it possible to cover a wider range of viscosity than hitherto. It is found that the viscosity of these substances increases enormously rapidly under pressure, much more rapidly than exponentially. Increases of viscosity of 10^7 fold have been produced by a pressure of 10,000 kg/cm^2. These results emphasize a conclusion previously reached—namely, that under the pressures in the crust the viscosity of liquids of complicated molecules must be so much increased that the substance behaves like a stiff glass.

A study has also been made with new apparatus of the effect of pressure on the melting point of these substances. New apparatus was necessary because the melting points lie in such a range that there is no suitable pressure-transmitting liquid. A study of the effect of pressure on the melting point appeared desirable because previous experiments left open the possibility that these substances might constitute an exception to the generalization made from all other experiments—namely, that there is no critical point between solid and liquid under pressure nor a maximum melting temperature, but the melting curve rises to indefinitely high temperatures under indefinitely increasing pressure. The new experiments showed that these substances constitute no exception. The phenomena which indicated the desirability of the experiments are a consequence of the enormous effect of pressure on viscosity.

The fact that there is no maximum melting temperature nor critical point does not mean that any liquid can be made to freeze at any temperature, no matter how high, by applying sufficient pressure. It is quite possible that freezing may be permanently suppressed and the liquid remain permanently in the subcooled noncrystalline condition because of the effect of pressure on the viscosity of the liquid in making impossible the formation of nuclei of the solid phase. Several examples of this have been found. It is a consequence of the effect of pressure on viscosity that there may be an optimum pressure for crystallization. A consequence of the existence of an optimum pressure is that on occasion freezing may occur when pressure on a liquid is *released*, something which is thermodynamically impossible under reversible conditions.

A statistical study of the effect of pressure on many polymorphic transitions indicates that the probability is that the crystalline form of any substance with at all complicated molecules is not the form with which we are familiar in the laboratory. This is illustrated by the conclusion reached by Birch in a paper to the National Academy of Sciences at its 1950 fall meeting to the effect that the seismic data seemed to demand a material deep in the crust with properties corresponding to no known material. The simplest inference is to a pressure form of some known substance. Birch suggested olivine. Experiments made with new apparatus on the shearing strength of olivine to 100,000 kg/cm^2 have shown a definite although not striking discontinuity in the shearing strength of olivine at 85,000 kg/cm^2. Until measurements are made of the volume in this pressure range there is no *certainty* that this discontinuity in shearing strength is of geological importance. The importance of polymorphic transitions is increased by the possibility that there may be electronic transitions as well as lattice transitions induced by pressure. Such transi-

* An expanded abstract of the address as given.

tions have beem observed experimentally in caesium and cerium. The volume changes associated with such transitions may be large. The possibility is also to be kept in mind that under pressure there may be new modifications stable only under a combination of shearing stress with pressure. An example of such a transition produced in the laboratory is probably metallic lithium.

The importance of a polymorphic transition for geology would be expected to be greater, other things being equal, the greater the associated change of volume. But there may be other effects than those due to change of volume. During the process of the transition it is to be expected that any inequalities of stress would be wiped out as the molecules rearrange themselves. It follows that a transition, even with so small a volume change as to be otherwise indetectible, might be an effective mechanism for isostatic adjustment by release of accumulated stresses. Experience in the laboratory indicates that there are a great many transitions with small volume changes. Statistically, the number of examples of transitions is inversely as the volume change up to the limit of measurement, which is about 0.01 per cent on the volume.

Experiments have shown the enormous effect of pressure in increasing the ductility of ordinary materials. This means that under high pressures the phenomena of plastic flow will play an increasingly important part. It is not usually realized how vitally the phenomena of plasticity in a solid differ from the phenomena of viscous flow in a liquid. In an "ideal" plastic solid the velocity of plastic yield is independent of the force; in a viscous fluid the velocity of viscous flow is proportional to the force. In the laboratory under ordinary conditions of testing the "ideal" plastic state is approximated to only in a very narrow range of deformations near the initial yield point. But at high pressures and very large deformations the ideal state is found to prevail over an indefinitely wide range of deformation. Parts of the crust of the earth may therefore be presumed to be in the ideally plastic state, and it would seem that more attention than appears to have been given up to now should be given to examining the mathematical consequences of the presence of such material in the crust. It is obvious that the boundary conditions will play a quite different role in a plastic earth than in a viscous earth.

Experimental knowledge of the properties of matter in the plastic state is meager. It has recently been found that some materials show in the plastic state near the fracture point in simple compression a reversible change of volume in the direction opposite that produced by the elastic action of the stresses. This raises the question of whether the elastic constants in general for small deformations, and the cubic compressibility in particular, of matter while in the process of undergoing plastic deformation are the same as the constants in the ordinary elastic range. This question is evidently pertinent to the problems of seismology. It would seem that an answer would not be difficult to obtain by the application of dynamic methods.

Plastic flow as ordinarily studied involves flow in three dimensions, as when a cylinder is shortened under a longitudinal load. By proper loading and constraints the flow may be made two- instead of three-dimensional. Such flow must be a closer approximation to much of the flow in the crust of the earth than three-dimensional flow. Recent studies made in the laboratory of such two-dimensional flow indicate certain qualitative differences as compared with three-dimensional flow. Strain hardening proceeds much more slowly in two-dimensional than in three-dimensional flow. It is much easier for instabilities to arise in two-dimensional flow. Mathematical singularities can easily arise in two-dimensional flow, as when a cube is sheared by tangential forces on two opposite faces. Under these conditions there is a mathematical singularity along the edge of the sheared face. Such mathematical singularities are accompanied by infinite stresses, the practical result of which is plastic flow. It would seem that plastic flow arising from mathematical singularities may be common in the crust. Every discontinuity in the elastic constants, as in the grains of one material embedded in a matrix of another, is a potential mathematical singularity.

Large plastic flow is accompanied by time effects which are not noticeable for small deformations. For instance, if an additional load

is applied to a material which has experienced a large deformation, yielding to the extra force may not begin at once, but may slowly build up in time to a maximum and then drop off again. Or the yielding may be spasmodic and jerky, and consist of long periods of quiescence punctuated irregularly with short intervals of rapid yield. The application to earthquakes is obvious.

Large plastic flow may be accompanied by very great strain hardening and increase of strength. For some substances the increase is linear in the strain ("natural" strain) and continues over the entire range of strain accessible to measurement. An expected consequence is that it would be extremely difficult to compress a powdered material down to the density of the homogeneous material because of the great strength developed at the corners of the small crevices by the unlimited strain hardening. This corresponds to laboratory experience.

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BEARING OF SOME BIOLOGICAL DATA ON GEOLOGY

By ERNST MAYR

(Address before The Geological Society of America, Sixty-third Annual Meeting)

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INTRODUCTION

When your Society honored me with an invitation to speak on this occasion, it was suggested that I treat some topic dealing with the relations between geology and biology. This at once brings up the question: What relations are there between the two sciences? Geology, by and large, consists of two major branches. One of these, Physical Geology, has no obvious contacts with biology, although we biologists derive much important information from the fields of geophysics and geomorphology. Much closer is the contact of Historical Geology with biology. Just how important this contact is to geology can be demonstrated by the following facts. Although the earth is some 3000 or 3500 million years old, little is known about the first 85 per cent of the time. Nearly all of our more detailed knowledge of the history of the earth is confined to the last 15 per cent or 500 million years—namely, that period which is documented by the fossil record.

When it comes to stating what part of biology is in contact with geology, the answer is a little more difficult. Biology is a large array of subjects that may indeed appear somewhat bewildering to an outsider. There is anatomy, morphology, histology, cytology, physiology, genetics, ecology, taxonomy—to mention only some.

However, it is possible to organize and classify all these subjects into two or three major categories. There are first the purely descriptive branches of biology which ask the question: What is there? The remainder of the field can be sorted out rather neatly by asking two other questions. The first of these deals with function: How does a given organism or a structure of an organism function? How does a chromosome function or a cell or the brain of a courting male bird of paradise? This large field of functional biology has rather little contact with geology. This operational approach to biology has yielded so many spectacular results in a causal analysis of biological phenomena that to many physiologists and biochemists this branch of biology has become virtually synonymous with biology as a whole. This is a short-sighted attitude because this functional approach leaves a great deal of the organic world unexplained.

A historical approach is needed to solve these remaining questions. We must ask the question: How did a particular structure, or organism, or group of organisms, or fauna evolve? These were the great questions in the passing days of the eighteenth century and the first half of the nineteenth century. Eventually this complex of questions was organized into a meaningful science by the theory of evolution. The whole field has therefore also been referred to as "evolutionary biology." Although the main outlines of this field and of our modern concepts were already laid down in Darwin's classic on the origin of species in 1859, there have been many advances since that date; in

fact, the field is still actively moving ahead. It is this branch of biology—since it deals with historical things—which has contact with Historical Geology. I shall discuss some of the problems and recent developments of evolutionary biology insofar as they appear to be of interest to the geologist.

It would be most misleading to imply that the contributions are one-sidedly those of biology to geology. On the contrary, geology has played a major part in the development of evolutionary thought. What a staunch ally of biology geology has been in the fight for recognition of evolution was emphasized by Darwin himself throughout his life. He stated on the occasion of the death of Sir Charles Lyell: "I never forget that almost everything which I have done in science I owe to the study of his great work." Lyell was a uniformitarian—that is, he believed in the slow transformation of the surface of the earth. In fact, his famous *Principles of geology* was written in 1830 as a direct answer to the cataclysmic theories of Cuvier propounded in the *Discours sur les Revolutions* etc., which was published five years earlier in 1825. So close, indeed, is the contact between geology and biology that Lyell in turn leaned heavily on the publications of Lamarck, one of the great early evolutionists. It is thus evident that both sciences are mutually indebted to each other, and that for at least 150 years there has been a fairly even give and take between them. In view of this close contact, it is necessary that the evolutionary biologist understands the principles and major findings of historical geology and that, in turn, the geologist be familiar with the corresponding elements of evolutionary biology.

The two subdivisions of evolutionary biology that are in specially close contact with geology are biogeography and paleontology. As far as paleontology is concerned, there has been a rather futile argument in recent years as to whether it is a branch of geology or of biology. The answer, of course, is that it is either both or neither. Like so many borderline sciences it overlaps both adjacent fields and binds them together. Fossils are the remainders of organisms that were once alive. The geologist who uses fossils for the establishment of geological sequences, or for the reconstruction of former climatic conditions, or other past events, must understand the biology of living organisms if he wants to interpret his material properly.

Anyone who has followed recent paleontological literature will realize what progress has been made in this field in recent decades. Not being a paleontologist I shall not enter into a detailed discussion of some of the more acute problems in this field. I will mention, however, three important principles, the neglect of which in the past has led to serious error.

The first is that every species is a variable population in which no two individuals are exactly alike. This principle may seem to us almost a platitude, but its antithesis, the typological concept of taxonomic categories, has been held rather widely until recent times and has been the basis of many untenable evolutionary theories. The replacement of the typological concept of natural populations and phenomena by a statistical concept has been one of the most revolutionary changes in biological thought.

The second important principle is that the composition of a local fauna and of the ecological communities of which it is composed are determined by the conditions of the environment. We take it for granted that at the present time we find different faunas in the Amazon forest, in the prairies of the Middle West, in the deserts of central Asia, or in the tundras of the far north. Even locally we find different animal communities in deciduous forest, in marshes, and in the sand dunes of the coast. In the oceans, also, different faunas are found along the rocky coasts of Maine, the sandy beaches of the Carolinas, or the mud flats of the Mississippi delta. All these are contemporary. We must remember this when we want to interpret different faunas of the past. They may well have been contemporary but lived in different environments.

The third principle also is an ecological one. The tremendous importance of the environment, for individual species as well as for entire faunas, has long been appreciated. Under environment in this connection we usually understand climate, topography, and, in particular, the distribution of the water and land on the continental shelves. However, we are only now beginning to realize that the living environ-

ment was as great or even a greater evolutionary factor than the inorganic one. The living or biotic environment consists not only of the plant cover of the earth, but also of competitors, of predators and other enemies, of parasites, and disease organisms. The disease factor is of incalculable potential importance in the decline and extinction of whole fossil families, as Haldane has indicated.

Time does not permit me to discuss in more detail these principles as well as many others recently dealt with by Simpson, Rensch, and other students of evolution. This much is certain, however—that the ever-increasing accumulation of fossil data, together with their interpretation in the light of the newly developed biological principles, will in due time lead to a considerable modification of our views on the past history of life on the earth.

These advances are of interest not only to the biologist but also to the geologist, as will be evident from a discussion of the relations of geology to another branch of biology—namely, biogeography.

Biogeography

Biogeography deals with the distribution of animals and plants. Such distribution is in part caused by factors of the contemporary environment and in part by historical factors, as for instance the former availability of land bridges. It is in this connection that a study of the distribution of present and former faunas and floras becomes exceedingly important to the geologist. The reason is the following:

It is one of the objects of geology to determine the historical changes in the outline and the position of our continents and oceans. It is of the greatest importance to know whether the continental masses and oceanic basins have always existed and whether they always had more or less the same relative position. It cannot be decided on the basis of purely geological evidence whether or not South America and Africa were formerly joined together and subsequently drifted apart. Even if we should have the same sequence of strata on both continents—let us say, red sands followed by shale, followed by limestone—, such a sequence might have been caused by similar climatic and tectonic conditions in rather far distant continents. However, if it can be shown that the faunas of two continents are essentially the same, it can be postulated with a good deal of assurance that there must have been a close land connection, because great faunal similarity is found today only between regions that are broadly connected by land bridges. In order to determine the location of former land bridges an analysis of formerly existing faunas is necessary.

Before discussing the results of such studies, it is necessary to say a word or two on methods. I have a feeling that some geologists are a little suspicious of much of the modern taxonomic work in paleontology. In particular they resent, with good reason, the vast multiplication of names, owing to much splitting of genera and species. These nomenclatorial byproducts of some of the recent work are regrettable, but they are not an inevitable result of a highly desirable trend toward a finer analysis. On the contrary, such splitting is often the consequence of a typological attitude and of a neglect of biological systematics. It is possible to make a detailed analysis of extinct animal populations without too great a burdening of the nomenclature. I will explain why such a more detailed and precise analysis is important.

Fossil animals or plants are used by the geologist as biological indicators of past events or conditions. They are used as scales or measuring tools to solve such problems as the sequence of geological periods, the length of geological periods, the nature of the deposits (whether terrestrial, fresh-water, or marine), the prevailing climate at the time of the deposit, and similar questions. The reliability of the answer to these questions depends on the accuracy of the utilized biological indicators. Anyone who has ever done experimental work in a chemical or physical laboratory knows that the chemist or physicist is constantly endeavoring to improve the accuracy and reliability of his measuring devices and to eliminate one after the other of the many potential sources of error.

The geologist, who uses animals or plants as indicators of past events or conditions, should realize that he too has the obligation to devote attention to a refinement of his measuring devices, some of which, as stated, consist of the fossil evidence. This is even more necessary

because biological data are subject to many more sources of variation than data drawn from the inorganic world. This is the principal practical reason for the ever-increasing precision of

less birds to which the emu, the cassowary, the ostrich, and the South American rheas belong, are found in all the southern continents (Fig. 1). This has been cited as proof for the former

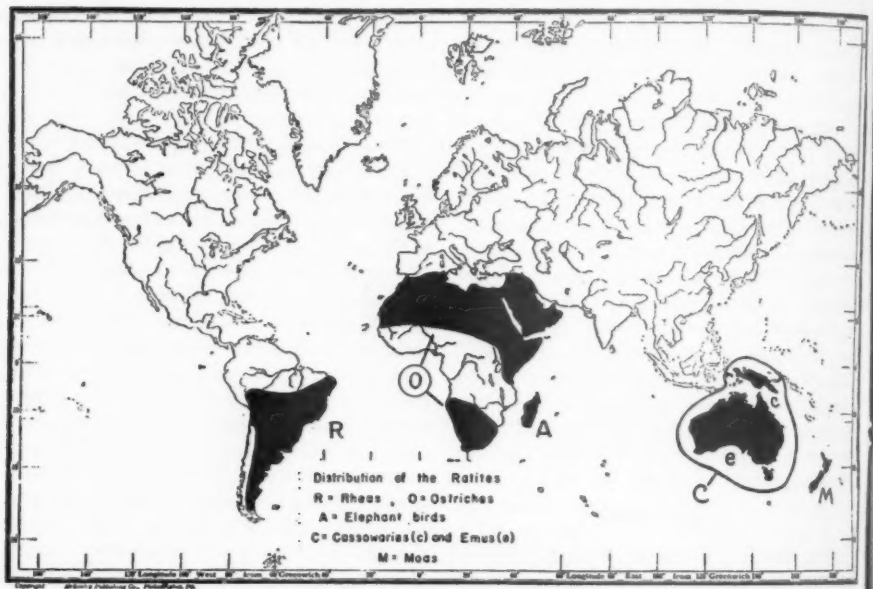

FIGURE 1.—DISCONTINUOUS DISTRIBUTION OF THE RATITES (LARGE FLIGHTLESS BIRDS) IN THE SOUTHERN HEMISPHERE

This distribution pattern cannot be cited as evidence for a former connection of these land masses since the five orders of Ratites are apparently not related to each other.

the paleontological taxonomic analysis. Many improvements in recent years of our knowledge of past events on the earth are due to elimination of biological errors owing to corrections of the systematics of living or extinct organisms.

Let us now go back to the field of biogeography and see how the elimination of taxonomic and other errors has led to scientific advances.

To begin with a concrete case. There are quite a few among the older biogeographers who have insisted that a trans-Atlantic land bridge between South America and Africa must have existed at least through part of the Tertiary. What is some of the evidence and why is it misleading?

The principal sources of error have been the following:

1. An incorrect determination of relationships. For instance, the ratites, the large flightless birds to which the emu, the cassowary, the ostrich, and the South American rheas belong, are found in all the southern continents (Fig. 1). This has been cited as proof for the former connection of these continents because how else could these flightless birds have possibly reached the now disconnected areas, particularly since these birds are now absent from the temperate zone of the Northern Hemisphere? The solution is that modern investigations have confirmed Fürbringer's views expressed as many as sixty years ago that the ratites are a polyphyletic group of unrelated, secondarily flightless birds and cannot be cited as evidence for a former connection of southern continents.

Let us consider just one other example: Simpson analyzed the Tertiary mammal fauna of South America and showed that all indigenous groups of mammals had reached South America from the north with the apparent exception of the hystricomorph rodents, which appeared to be closely related to the Old World porcupine but without obvious Tertiary an-

cestors in North America. This puzzle was recently solved by Wood (1950) who showed that again, as in the case of the ratites, we are dealing with a case of convergence and that both the South American and the African porcupines have evolved in a parallel manner from a group of primitive rodents, the Paramyidae, which, in the lower Eocene, had a wide distribution in the Northern Hemisphere.

Recent taxonomic research has cleared many such polyphyletic categories, resulting in a better understanding of former distributions (*e.g.*, Simpson, 1940).

2. A second source of error is the underestimation of the faculty of dispersal. The map (Fig. 2) of a species of tree ducks, *Dendrocygna*, showing that the same species occurs in Africa and South America but is absent in the north, would seem to indicate a recent connection of the two continents. A further study of the case, however, indicates clearly that this fresh-water duck must have had the ability to cross the ocean occasionally and to establish in this manner the otherwise irrational distribution. Many similar cases have been cited as proof for a trans-Atlantic land bridge.

The agents of dispersal are manifold, and their activities often quite unbelievable. The occurrence of fairly rich faunas and floras on the volcanic islands of the Pacific basin from Samoa to Hawaii and the Tuamotus indicates that occasional tropical hurricanes and migrating birds, if given enough millions of years, can produce wonders of dispersal.

As soon as this is clearly realized, most of the land bridges become unnecessary that were postulated by biogeographers of former generations. A discontinuity in the range of a species or other taxonomic group is by no means proof for a former direct contact of the areas in question. For instance, there are many highland forms of northern relationship in the highlands of Guatemala which must have jumped across the tropical lowlands of the Isthmus of Tehuantepec where climatic conditions were never suitable for them, not even during the coldest period of the Pleistocene. The existing faunas of the West Indian islands do not require a land connection with the American mainland

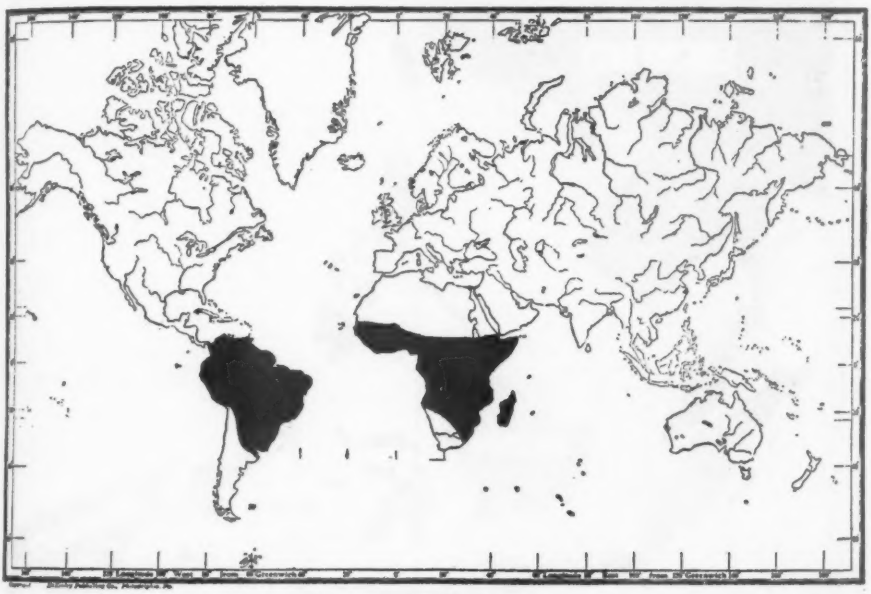

FIGURE 2.—RANGE OF THE FRESH-WATER DUCK *Dendrocygna viduata* ON BOTH SIDES OF THE ATLANTIC

nor does the Tertiary fauna of Madagascar require connection with Africa, as Simpson has shown so clearly.

ing maps. This is particularly true if the current distribution is taken as the basis and the former distribution reconstructed by inference.

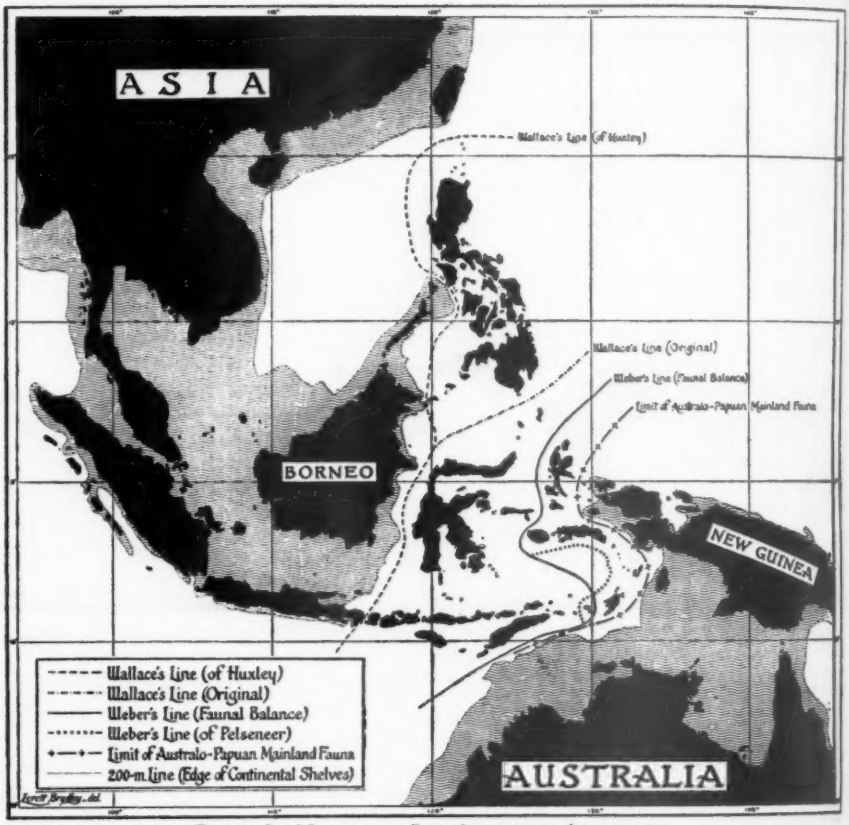

FIGURE 3.—MAP OF THE INDO-AUSTRALIAN ARCHIPELAGO

The lines drawn by the various zoogeographers ("Wallace's Line, Weber's Line, etc.") to separate the Indian from the Australian region run from north to south. There is a striking difference between a western and an eastern fauna.

I would like to emphasize particularly the need for a more penetrating study of dispersal. It will solve many puzzles of the biogeographer and paleontologist. It will settle many controversies of long standing. For instance, if a geologist should ask, first a student of mammals, then one of insects, then one of conifers or of grasses, to reconstruct a series of paleogeographic maps on the basis of the present and former distribution of their respective groups he is likely to receive four rather different-look-

Let us take, for example, the Indo-Australian region of today. The students of mammals, fishes, birds, and other groups draw a sharp line roughly from north to south through the Malay Archipelago, a line which separates the Indian fauna in the west from the Australian fauna in the east (Fig. 3). The two faunas are exceedingly well defined even though there is a certain amount of overlap in the insular zone of contact. If you ask a botanist to draw a phytogeographic map of the same region, it

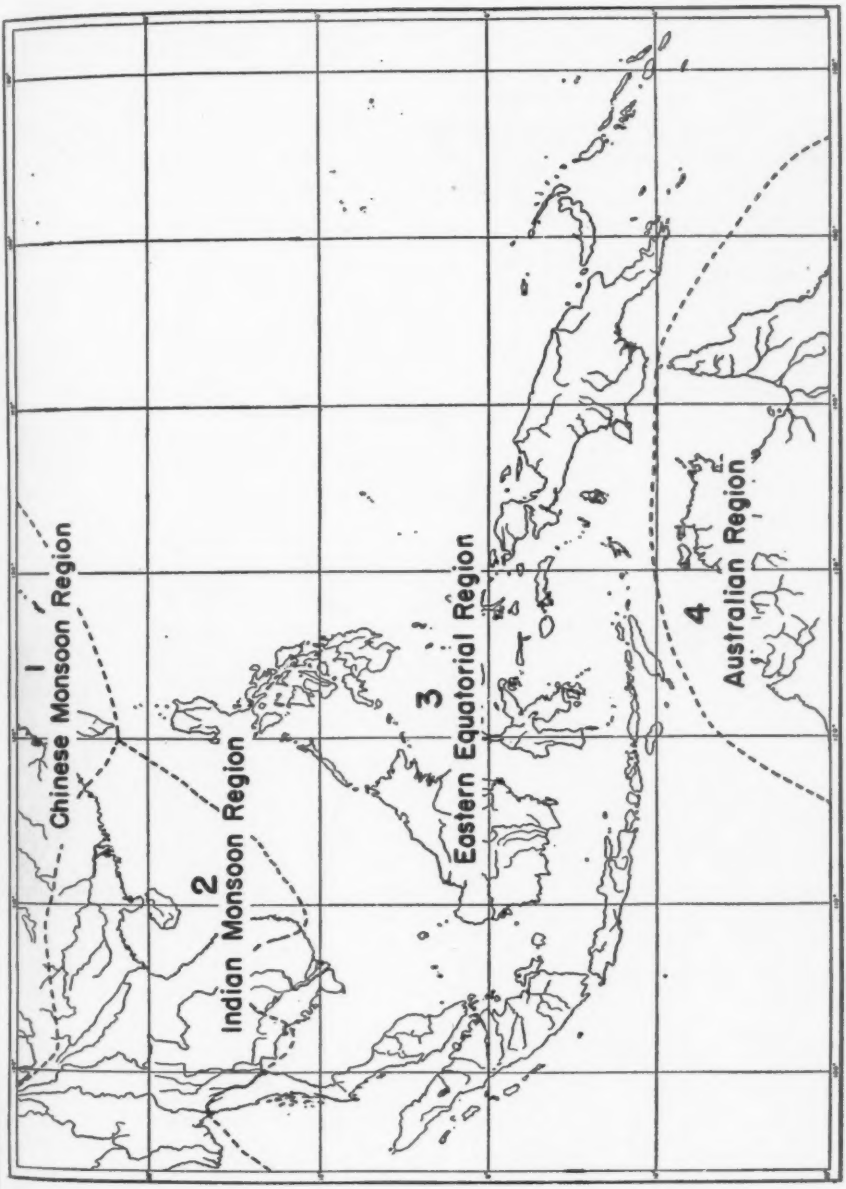

FIGURE 4.—PHYTOGEOGRAPHIC REGIONS IN THE INDO-AUSTRALIAN ARCHIPELAGO
After Shaw, 1943.

will be drastically different (Fig. 4). The phytogeographer may combine the whole tropical belt into a single region and draw a line from east to west across tropical northern Australia

Guinea, and in the eucalyptus scrub of the interior of arid Australia. The sharp division between the Indian and the Australian fauna, as shown by birds, mammals, and other groups

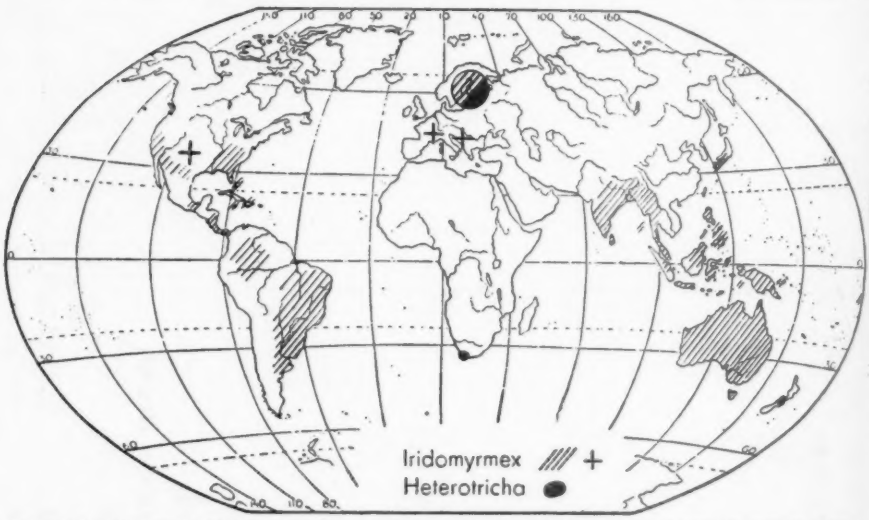

FIGURE 5.—WIDELY DISCONTINUOUS RECENT RANGES OF GENERA RECORDED IN THE OLIGOCENE BALTIC AMBER

\+ = Additional early Tertiary range of *Iridomyrmex*. From Ander (1942).

separating a Malayan flora from an Australian one.

These two kinds of maps suggest exceedingly different geological histories during the Tertiary, which is, of course, impossible. What is the explanation for this discrepancy? The primary factor is that plants have very high dispersal facilities but are ecologically very demanding. They spread rather easily through the ecologically fairly uniform tropical belt from Malaya to New Guinea and the Solomon Islands but have great difficulty in becoming established in the rather arid subtropical zone of Australia. Animals, on the other hand, and warm-blooded vertebrates, in particular, have much greater dispersal difficulties, but once they have jumped an oceanic barrier they have apparently less difficulty in adjusting themselves to a new environment. The result is that we have many genera of birds and mammals that are equally at home in the primeval rain forest of the lowlands of New Guinea, in the moss forest of the high mountains of New

of animals, is an indication of the long-standing separation of the Asiatic Sunda Shelf and the Australian Sahul Shelf. The similarity of the floras in the tropical belt is an indication of the similarity of climatic condition.

It would be quite understandable if a geologist would draw very different conclusions from zoogeographical and phytogeographical maps of the Indo-Australian archipelago. A valid analysis depends in this case, as in so many others, on the proper evaluation of the biological evidence. It will solve many of the discrepancies that now plague the paleontologist.

Let me cite two other sources of error. The first is the explicit or implicit assumption that the currently found distribution patterns of species—or of higher taxonomic categories—are as old as the species themselves. This assumption is completely wrong. Distribution patterns, as found today, are the joint product of present and of past conditions. The ranges of species shift from year to year in response to changes in climatic conditions. At the peak

of the Pleistocene glaciation some arctic animals ranged as far south as the southern United States, and, as everyone knows, tropical animals ranged much farther north during the Mesozoic and early Tertiary. Changes in climate result in large shifts of geographical ranges.

Today, the camel family is found only in South America and central Asia. This does not mean that there once must have been a trans-Pacific land bridge between the two areas, as was seriously suggested. The real solution is that North America at one time was the center of distribution of the camel family as proven by the fossil record. Maps (Fig. 5) of the present distribution of insect groups found in the Oligocene in the Baltic amber of northern Europe also show how misleading current distribution patterns can be.

One other source of error is the neglect of geological time. It would be nonsense to cite the occurrence of Mesozoic plants and of recent birds as evidence for a former continental connection of Greenland and northern Europe. This argument, of course, works both ways. If, for instance, it has been demonstrated conclusively that all the available evidence contradicts a continental connection between Africa and South America during the Tertiary, this does not mean that it is thereby also proven that such a connection could not have possibly existed in the early Mesozoic or in the Paleozoic. The possibility of faunal and continental connections must be studied separately for each geological period.

Results

Let us now see how these newly sharpened tools can be applied toward the solution of specific geological problems. With regard to continental connections the geologist wants to know primarily three facts:

(1) Was there such a connection?

(2) When did it exist and for how long a period?

(3) What were the ecological and climatic conditions in the connecting area?

In a thorough analysis these questions should be asked for each of the postulated former land connections.

The adherents of continental drift postulate, for instance, that Australia was in contact with Antarctica and South America during the early part of the Cenozoic while during the same period it was widely separated from Asia.

They believe that Australia gradually drifted toward Asia during the later Tertiary and finally crashed into the festoons of islands situated along the outer edge of the Asiatic shelf. If this picture were correct, we would expect Australia to have a fauna consisting of two drastically different elements: an old one with close South American relationships and an exceedingly new one of strictly Malayan character. On the other hand, if the orthodox geological picture is correct that Australia has not materially changed its geographical location since, at least, the latter part of the Mesozoic and was always separated from Asia by an oceanic region more or less studded with islands, then we would expect an Australian

Table 1.—Analysis of Recent Australian Bird Fauna

Category	Number of Units	Estimated Time of Arrival
A—Autochthonous families and subfamilies of unknown relationship but not South American	15	? Eocene-(Oligocene)
B—Autochthonous families and subfamilies of Asiatic origin	8	? (Eocene)-Oligocene-(Miocene)
C—Endemic genera with Asiatic relatives	*ca.* 40	Miocene-Pliocene
D—Endemic species of Asiatic origin	62	Pliocene, early Pleistocene
E—Asiatic species, some with endemic Australian subspecies	40	Middle Pleistocene to Recent

fauna of ultimately Asiatic origin but consisting of successive layers of younger and younger faunal elements which jumped the island chains between Asia and Australia during successive stages of the Mesozoic and Tertiary.

I have analyzed the bird fauna of Australia, and the following picture emerges (Table 1). There is no evidence of any relationship with the South American bird fauna. There is no evidence of only two faunal elements, an old and a very new one, but rather the fauna of Australia is indeed composed of successive layers of elements that show closer and closer relationship to the Asiatic fauna. The zoogeographic analysis of the recent bird fauna of Australia thus confirms decisively the orthodox findings of geology.

Conclusions

I am at the end of the presentation of my material. I sincerely hope that my discussions have made it evident how many-sided the contacts are between biology and geology. The geologist must always be aware of the fact that he uses biological findings mainly as indicators of past geological events and conditions. However, he is seriously in danger of misusing these indicators if he does not realize the complexity of the biological material. Biologists have made rapid progress during recent years in uncovering the various sources of error in the use of biological material. We hope that by eliminating these errors biological indicators will become increasingly useful to the geologist in helping him to solve at least some of his many problems.

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